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MEASURING GEOLOGIC TIME BY THE STRONTIUM METHOD

BY LOUIS AHRENS

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ABSTRACT

The strontium method for measuring geological age has been examined in each of its aspects, and the general conclusion is made that evidence is now sufficient to show this method as capable of providing reliable age measurements.

Of the total strontium in the earth's crust, 0.5–1.0% is considered to be radiogenic, almost all of which is concentrated in potassium and cesium minerals. Lepidolite contains the highest proportion of radiogenic strontium, in which mineral it usually predominates completely. Amazonite, pollucite, hydrothermal pegmatitic microcline, zinnwaldite, some specimens of lithium-rich muscovite, and probably also rhodizite and lithium-rich biotite commonly contain a high proportion of radiogenic strontium; age measurements may be made on most of these minerals, if not all. Age determinations on other minerals are also possible, and the strontium method could probably be extended to granite biotites, which would increase its scope very considerably. With lepidolite, a reasonably reliable age can be obtained in most instances without an isotope analysis, although one is desirable. An isotope analysis is imperative for age determinations on all other minerals. For a quantitative isotope analysis of strontium, the mass spectrograph is superior in sensitivity and accuracy to other methods; a quantitative analysis can be made on as little as 0.3 mg. of strontium salt, and a conveniently low proportion of radiogenic strontium can be determined in the presence of an excess of common strontium.

In all, 32 strontium age determinations are known, 30 of which have been made according to a spectrochemical procedure outlined in this paper. The procedure is rapid—10–15 determinations (without isotope analyses) may be made in quadruplicate in 3 days—but lacks precision, since age reproducibility is usually only within about ±10–15%, even when carried out in quadruplicate. A more accurate procedure for determining Sr/Rb ratios is urgently needed to increase the usefulness of the strontium method; investigation of other spectrochemical methods is recommended.

Where comparisons were possible, strontium ages tally reasonably well with lead and helium (magnetite) ages, with few exceptions.

The strontium method should be of particular value in pre-Cambrian time and is probably superior to the lead and helium methods for dating very ancient rocks. The extreme antiquity of pegmatites from southeast Manitoba has been established beyond reasonable doubt; their age (2100×10^6 years) is considered to be greater

than any other region on the earth's crust on which sufficient data are available. The method is, however, apparently not so suitable for dating relatively young specimens, and even under the most favourable circumstances, that is, if lepidolite is available, it seems unlikely that an age of less than 50×10^6 years could be measured successfully.

Lead, helium, and strontium methods have been compared.

INTRODUCTION

Some 10 years ago, Goldschmidt (1937a, p. 140) and Hahn and Walling (1938) suggested that the natural beta decay of rubidium to strontium could be utilized for determining the geological age of a mineral. During the decade since, considerable investigation has been carried out on the use of the strontium method, and some speculation, a fair proportion sounding a note of pessimism, has been made on the scope and limitations of the method.

In this paper the strontium age method is examined in some detail in each of its aspects and several new age determinations and other data are given, in addition to an account of some recent investigations on methods of analysis. It is hoped that after perusing this paper the interested reader will have cause to feel that within a reasonable period this elegant method for measuring geological time should provide reliable and accurate age data, particularly on pre-Cambrian time.

It is hoped further that active interest in this method will be stimulated: in each of the several aspects of the strontium method, only preliminary investigation has been made, and there is ample room for a tremendous amount of research.

A preliminary report of a portion of the work described in this paper has been given by Ahrens (1947a).

ACKNOWLEDGMENTS

Much of the earlier research described in this paper was carried out at the Government Metallurgical Laboratory, University of the Witwatersrand, Johannesburg, and thanks are due to the Director, Prof. L. Taverner, for permission to use laboratory equipment for spare time research. Mr. J. Levin, also of the Government Metallurgical Laboratory, kindly prepared some very pure specimens of lepidolite by a mineral dressing procedure. Grateful acknowledgment is also made to staff members of the Geology and Physics departments of the Massachusetts Institute of Technology where most of the later research was carried out. Thanks are extended to Prof. W. J. Mead and Prof. H. W. Fairbairn of the Dept. of Geology, for generously providing facilities and all necessary operational requirements; and to Dean G. R. Harrison and Dr. R. McNally (Jr.) of the Physics Department, also for providing facilities and for making useful suggestions for the recording of isotope effects. It is a pleasure also, to acknowledge the ready co-operation of several persons in South Africa, America, and elsewhere in providing specimens.

A major portion of the research described in this paper was carried out when I was a Research Fellow of the South African Council for Scientific and Industrial Research; grateful acknowledgment is made to this Council for having provided the necessary opportunity for this research.

NATURAL RADIOACTIVE DECAY OF RUBIDIUM

Like its alkali metal homologue, potassium, rubidium is beta active, and it decays into a stable isotope of strontium. Rubidium consists of two isotopes, Rb^{87} and Rb^{85}, the relative abundances of which are 27.2 and 72.8%, respectively; according to Hahn, Strassmann, and Walling (1937) and Hemmendinger and Smythe (1937), Rb^{87} alone is radioactive, whence the above transmutation may be written as:

$$Rb^{87} \rightarrow Sr^{87} + \textit{beta particle}$$

For convenience, radiogenic strontium will be written thus, Sr^{87*}, so as not to confuse ordinary Sr^{87} found in common strontium.

The radioactive disintegration of any radioactive element may be expressed as

$$N = N_0 e^{-\lambda t} \quad \ldots\ldots (1)$$

where N_0 is the number of atoms originally present and N the number left after a time, t, has elapsed; λ is the decay constant. In the above equation, N_0 and N refer to atoms of Rb^{87} and λ its decay constant: the period of half life, T, is equal to $0.693/\lambda$.

Since, as will be shown later, the half life is very long in comparison to geologic periods, only an insignificant proportion of rubidium will have decayed even in the most ancient minerals (Fig. 1), and consequently one may regard the quantity of rubidium in a mineral as having remained constant throughout its lifetime. Equation (1) can then be simplified and written as follows,

$$N = N_0 - \lambda t N,$$
$$N_0 - N = \lambda t N \quad \ldots\ldots (2)$$

$N_0 - N$ = the number of atoms of Rb^{87} that have decayed, which is equivalent to the number of strontium atoms that have formed, whence we may write,

$$\%Sr^{87*} = \%Rb^{87} \times t \times \lambda,$$

or

$$t\ (\text{age}) = \frac{\%Sr^{87*}}{\%Rb^{87}} \times 1/\lambda \quad \ldots\ldots (3)$$

Figure 2 compares the use of the simplified equation (3) with equation (1). For a given amount of Rb^{87}, calculations were made using these two equations, of the quantities of Sr^{87} generated at various times and Figure 2 shows a plot of per cent variation vs. time. The amount of deviation corresponds to the percentage deviation in the age measurement if the simplified equation is used; for most of geologic time the deviation is small, and the convenient simplified equation may therefore be used.

The accuracy of a determination of geologic age of a mineral by the strontium method will depend upon the accuracy with which λ, $\%Rb^{87}$, and $\%Sr^{87*}$ can be estimated. The experimental determination of Rb^{87} and Sr^{87*} is discussed later.

Two relatively early determinations of λ (Hahn and Rothenbach, 1919; Mühlhof, 1930) are equal to 9.2×10^{-12} yr.$^{-1}$ and 6.0×10^{-12} yr.$^{-1}$ respectively. A more recent and indirect determination is reported by Strassmann and Walling (1938). From lepidolite from the Silver Leaf mine, Southeast Manitoba, Canada, strontium

was extracted, determined quantitatively and analysed mass spectrographically. The analysis showed > 99.7% of the strontium to be radiogenic; the rubidium content of the lepidolite was also determined. The age of the lepidolite was con-

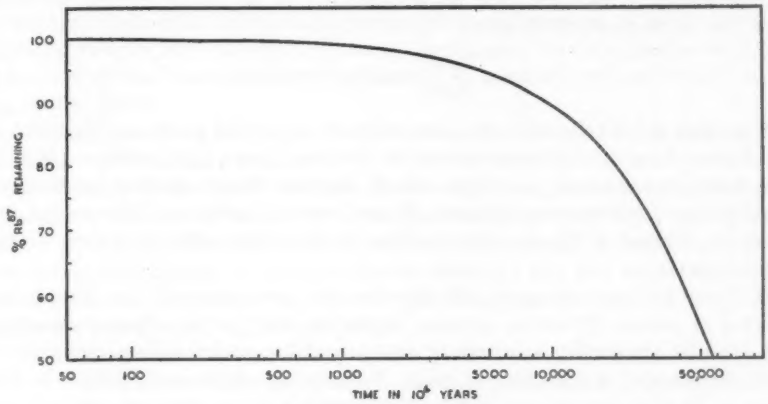


FIGURE 1.—*Decay of Rb^{87}*

Showing %Rb^{87} remaining after different periods have elapsed.

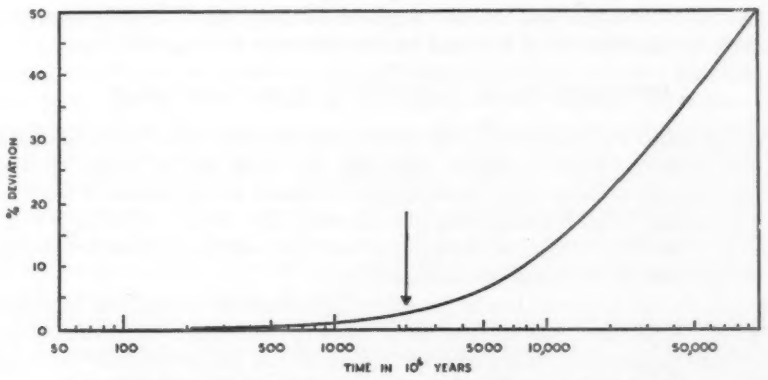


FIGURE 2.—*Percent deviation in computing ages by use of simplified equation*

(Arrow indicates age of oldest known mineral).

sidered to be equal to that of the near-by Huron Claim pegmatite which had been dated by lead age measurements. An age of 1975×10^6 years was used in this computation, and λ was calculated as 1.11×10^{-11} $year^{-1}$ and T as 6.3×10^{10} years. Unfortunately this age is somewhat approximate, and according to Goodman and Evans (1940, p. 284–287) the value of λ may be in error by as much as 30–40%.

The above value of λ has been used for all strontium age determinations so far reported, but does not appear to be in serious error because a new value of 5.81×10^{10} years for the half life of Rb^{87} has recently been determined by Eklund (1946)

using a direct method; this value is in reasonably good agreement with the indirect determination and has been used for all the age determinations reported in this paper.[1]

Substitution of the new value for λ ($0.693/5.8 \times 10^{10}$ years $= 1.195 \times 10^{-11}$ year$^{-1} = \lambda$) in equation (3) gives,

$$\text{Age} = \frac{\%Sr^{87*}}{\%Rb^{87}} \times 8.37 \times 10^{10} \text{ years} \quad \ldots\ldots(4)$$

The value of λ is believed to be known with an accuracy of about $\pm 15\%$, which is satisfactory for general purposes but not for work requiring a high precision. Within reasonable time however, according to Prof. Sanborne Brown (private communication) of the Department of Physics, Massachusetts Institute of Technology, an accuracy of about $\pm 5\%$ should be possible, which should suffice for almost all age requirements.

Rb^{87} and Sr^{87} are isobars, that is, they have the same mass (A), but differ in the number of protons (Z) in the nucleus. Rubidium has $Z = 37$, whereas strontium has $Z = 38$. According to a rule of nuclear stability, stable isobaric pairs which differ by one unit of Z should not exist. Very few exceptions are known. In this instance Rb^{87} is unstable, and Rb^{87} and Sr^{87} are thus not exceptional, although the rate of decay is very slow. Furthermore, nuclear theory indicates that the greater the difference in spin between parent and daughter nuclei, the slower the rate of decay. Rb^{87} has a spin of 3/2 and Sr^{87} a spin of 9/2: this large difference apparently accounts for the existence of Rb^{87} and its very slow rate of decay to Sr^{87}.

GEOCHEMISTRY OF RUBIDIUM IN IGNEOUS MINERALS

The abundance of rubidium in the earth's crust has been given by Goldschmidt (1937a) as about 0.03% by weight. Although this value may be subject to some revision because of more recent quantitative abundance measurements of rubidium in minerals and rocks, the magnitude is in all probability correct; rubidium is therefore not a relatively abundant element nor very rare, which of course is fortunate for the purposes of geological age measurement.

It may be well to review briefly the general geochemistry of rubidium in igneous minerals. Rubidium forms no minerals of its own, but tends to substitute for potassium in potassium minerals and for cesium in the cesium mineral, pollucite; rubidium may also be present in some specimens of beryl, but the nature of its location within this mineral is not clear. Since Rb^+ (radius = 1.49 A) is larger than K^+ (radius 1.33 A) for which ion it proxies during crystallization of minerals, rubidium tends to concentrate relative to potassium during the formation of potassium minerals, in the mother liquor fractions (residua), in accordance with the general geochemical rule (Goldschmidt, 1937b, p. 661) that when two ions of like charge but of slightly different size compete for a given lattice site within a growing crystal, the smaller ion will be accepted in preference because of its greater electrostatic attraction. This rule undoubtedly holds for the pair rubidium: potassium. Evidence

[1] Two separate determinations of the half life of Rb^{87} have just been reported by O. Haxel, F. G. Houtermans, and M. Kemmerich (Phys. Rev., vol. 74, p. 1886-1887, 1948). After considering both their determinations, these authors give $6.0 \pm 0.6 \times 10^{10}$ yr. as their best value: this value is in close agreement with the value used in this paper, and the use of the newer value would alter the strontium ages given only very slightly.

seems ample that rubidium concentrates relative to potassium in granite pegmatites, and that, in these pegmatites, rubidium is concentrated particularly in those minerals typical of late hydrothermal replacement phases, such as lepidolite and hydrothermal microcline.

Except in potassium minerals, pollucite, and some specimens of beryl, rubidium finds no suitable host minerals and cannot be squeezed into the smaller structure holes and replace ions such as Na^+ (radius 0.98 A) and Ca^{++} (radius 1.06 A) to any significant extent.

Minerals containing rubidium will now be discussed in approximate order of decreasing rubidium content.

LEPIDOLITE: Specimens of lipidolite from many areas have been analyzed and there seems to be no doubt that with the possible exception of the extremely rare mineral, rhodizite, discussed later, lepidolite is the mineral richest in rubidium. The mean Rb_2O content in lepidolite may be taken as 1.5%, and the extreme limits of variation are from 0.5–3.0% (Stevens, 1938; Berggren, 1940; Ahrens, 1945a give many newer analyses of rubidium in lepidolite).

Several minerals which contain appreciable quantities of rubidium but usually less than in lepidolite can be roughly grouped together: hydrothermal pegmatitic microcline, amazonite, pollucite, zinnwaldite and some specimens of lithium-rich muscovite and biotite. In the absence of further information, rhodizite may be grouped here as well.

HYDROTHERMAL PEGMATITIC MICROCLINE: Not many analyses are available, but it seems definite that rubidium is very enriched in hydrothermal microcline, compared with the more common primary pegmatitic microcline, by a factor which may be as high as 5–10. Since the formation of hydrothermal microcline is a much later phase than that of the primary microcline, the marked enrichment of rubidium relative to potassium in the former is an interesting corroboration of Goldschmidt's (1937b, p. 661) geochemical rule. Adamson (1942) has found a high rubidium content (frequently about 1.0%) in hydrothermal microcline from the Varuträsk pegmatite from Sweden; the author has examined some specimens of microcline from Dixon, New Mexico, which are considered to be hydrothermal, and found a marked enrichment of rubidium.

AMAZONITE: The green variety of microcline is also considered to be hydrothermal but is characteristic more of granite druses than of pegmatites. Several analyses of rubidium in amazonite have been reported by Goldschmidt *et al* (1934), Tolmacev and Filippov (1935), Ahrens (1945a) and Erämetsä *et al* (1941). The mean rubidium content is about 0.5% and varies between limits of 0.1–2.5%.

POLLUCITE: In polllucite, Rb^+ undoubtedly substitutes for Cs^+ (radius 1.69 A). The analyses of rubidium in several specimens of pollucite (Ahrens, 1947b) indicate that the abundance of rubidium in pollucite is very similar to that commonly found in amazonite. Pollucite is a rare mineral and its occurrence in pegmatite has been reported in only about a dozen localities.

ZINNWALDITE: Information on the concentration of rubidium in zinnwaldite is very meagre, but it is usually present in smaller amounts than in lepidolite and seemingly in the same magnitude as in the minerals just discussed.

LITHIUM-RICH MUSCOVITE: The variation of the rubidium content in moscovite is

extreme, 0.005–2.0% Rb_2O. Evidence seems sufficient to show in a general way, a sympathetic variation between lithium and rubidium in muscovite. Stevens and Schaller (1942) mention the enrichment of rubidium in muscovite of late formation in complex lithium pegmatites; in many specimens of muscovite so far analyzed, those specimens enriched in lithia are frequently enriched in rubidia (Ahrens, 1945a).

One cannot refer to the association of rubidium and lithium as one could, for example, to element pairs such as gallium: aluminum and potassium: rubidium. Li^+ and Rb^+ behave as two different species of ions; Li^+ is octahedral in co-ordination, whereas Rb^+ is of higher co-ordination and occupies a different lattice site. Since Li^+ carries a single charge, its electrostatic attraction is weaker than that of the other octahedral ions such as Fe^{++}, Mg^{++}, Al^{+++} (octahedral) for which ions it has to proxy, and it tends therefore to be rejected in earlier crystals and enriched in the residua. Because Rb^+ enriches also in residual liquors, an enrichment of both lithium and rubidium is found in the same mineral if suitable potassium host minerals which have sites for ions of six-fold (octahedral) co-ordination are available, such as mica but not feldspar.

It is difficult to give a mean Rb_2O content in lithium-rich muscovite, but a high proportion of specimens appear to contain 0.1–0.8%.

Pegmatitic Biotite: Biotite is interesting; very little information is available, but apparently, biotite from complex lithium pegmatites is likely to be enriched in rubidium, in contrast to biotite from granite and from simpler pegmatites. Hess ans Stevens (1937) report as much as 1.85% Rb_2O in pegmatitic biotite from King's Mountain, North Carolina. A spectrochemical analysis (1.4% Rb_2O) by the author of some biotite (not the same specimen) from the same locality verifies the magnitude of rubidium found in their analysis. Another spectrochemical analysis of biotite, Strickland Quarry, Connecticut, also gave a high rubidium content, 0.7% Rb_2O. A spectrochemical analysis of muscovite from the same pegmatite revealed only 0.02% Rb_2O. Both specimens of biotite were found to contain appreciable quantities of lithium. More analyses of rubidium in biotite are given by Stevens and Schaller (1942).

In contrast Nockolds and Mitchell (1948) find on an average only about 0.15% Rb_2O in biotite from granite.

Rhodizite: An analysis of a specimen of this extremely rare mineral from Madagascar, quoted by Lacroix (1922, p. 342) reports 2.3% Rb_2O. A specimen from Manjak, Madagascar was kindly placed at my disposal by Dr. W. F. Foshag of the U. S. National Museum, and after a microscopic examination which was necessary to identify the mineral, it was analyzed spectrochemically. Unfortunately the mineral produced rather a poor spectrum and, since a very small quantity of the rhodizite was available, an analysis had to be made on the only spectrum that could be photographed, the background of which was very heavy. Consequently, a relatively accurate measurement of the rubidium content, or of the strontium content, could not be made, but the magnitude should be correct in each case. Approximately 1.0% Rb_2O was found, thereby confirming that rhodizite is enriched in rubidium.

Rhodizite is found in pegmatite, but its occurrence is so rare that although probably suitable for making age measurements, its usefulness would be severely curtailed.

BERYL: Many specimens of beryl appear to be barren of rubidium, but the type commonly referred to as cesium beryl may contain appreciable quantities, about which however, very little is known. In a specimen of cesium beryl associated with the specimen of rhodizite referred to above, about 0.3% Rb_2O was found.

MICROCLINE (AND PERTHITE) OTHER THAN THE HYDROTHERMAL TYPES, ORTHOCLASE, GRANITE BIOTITE AND POTASH FELDSPAR, CELSIAN, SANIDINE, LEUCITE, AND PHLOGOPITE: In minerals of this group, rubidium is usually much less abundant than in the other minerals so far discussed, but except for common pegmatitic microcline, several analyses of which have been reported (Ahrens, 1945a), analytical data on these minerals is very scant. In common microcline the rubidium concentration varies from about 0.003–0.2%. The reader may refer to Nockolds and Mitchell (1948) for data on granite biotite; to Goldschmidt *et al.* (1934) for data on celsian, sanidine, leucite, and other minerals, and to Stevens and Schaller (1942), Erämetsä *et al.* (1941) and Ahrens (1945a) for analyses of phlogopite.

Most of the more important publications giving information on the abundance of rubidium in different minerals are cited and should suffice to give the reader enough information on the distribution of this element. The abundance of rubidium in rocks is discussed later with the possible application of the strontium method to rocks.

The abundance of rubidium in some of the minerals discussed is quite high and could be estimated. The question now arises: as the decay rate of rubidium is very slow are the quantities of radiogenic strontium generated even in the oldest minerals richest in rubidium sufficient to permit a reasonably accurate quantitative analysis? Assuming the mean Rb_2O content of lepidolite to be 1.5%, the following quantities of strontium will be generated after periods of,

2000×10^6 years	0.01% Sr,
1000×10^6 years	0.005% Sr,
500×10^6 years	0.0025% Sr,
250×10^6 years	0.0013% Sr,
100×10^6 years	0.0005% Sr.

These quantities of strontium are very small, but, as will be shown later, they can be estimated satisfactorily by a spectrochemical procedure.

GEOCHEMISTRY OF STRONTIUM IN IGNEOUS MINERALS

Another question arises—will the rubidium-rich minerals be sufficiently free from an excess of common strontium to permit age measurements? The abundance of strontium in the earth's crust, although not known with great accuracy, is approximately equal to that of rubidium. Of this total strontium, 0.5–1.0% is considered to be radiogenic (Ahrens, 1948a); consequently, 99.0–99.5% is common strontium, and contamination, potentially very great, will depend on how similar the geochemical characteristics of rubidium and strontium are in igneous minerals.

The geochemistry of strontium is discussed and abundance data given only as pertinent to the problem of geological age.

On a coarse scale, the geochemistry of rubidium is similar to that of strontium:

both elements form typical cations of large size (radius of Sr^{++} = 1.27 A) having about the same co-ordination in silicates, and hence both elements are typical of silicates. At first sight this might seem rather discouraging, but their 'fine' behaviours are appreciably different.

A distinct shift in the potassium: rubidium ratio is found during the fractional crystallization of minerals because of a difference of ionic sizes. The difference between the sizes of Rb^{+} and Sr^{++} is still greater, and furthermore Sr^{++} is doubly charged, indeed a fortunate circumstance for the determination of geological age. As a result, strontium tends to associate with calcium (radius Ca^{++} = 1.06 A) in calcium minerals rather than with potassium; Sr^{++} will substitute for K^{+}, but to a limited extent. However, because Sr^{++} is doubly charged and smaller than K^{+}, strontium in potassium minerals tends to concentrate in the minerals of earlier crystallization. This point has been made by Noll (1934) who has investigated the geochemistry of strontium in some detail. Theory would indicate, therefore, that rubidium and strontium are geochemically antipathetic during the formation of potassium minerals. Rubidium concentrates in minerals of late formation; strontium concentrates in the earlier fractionates. Experimental evidence substantiates this antipathetic behaviour.

LEPIDOLITE: Leipidolite is very probably the mineral freest from contamination by common strontium. The isotope composition of strontium in a few specimens of lepidolite has been examined quantitatively by Mattauch (1947) with results as follows (Pl. 1):

Quantitative Measurements (Mattauch, 1947)

Source	Proportion of radiogenic strontium
Southeast Manitoba (1)	99.7%
Southeast Manitoba (2)	99.6%
Southeast Manitoba (3)	98.2%
Varuträsk, Sweden	97.4%

A hyperfine structure analysis of certain strontium optical (emission) line spectra by Heyden and Kopfermann (1937) on strontium from specimen (1) above corroborated the mass spectrographic analysis.

Qualitative measurements (Brewer, 1937) of three specimens of lepidolite from Pala, California, and a specimen of zinnwaldite, detected only radiogenic strontium. Although qualitative, these results are of considerable value because the age (Nevadan) of the Californian batholith with which the pegmatites containing the lepidolite are associated is relatively recent, about 100×10^6 years, and the lepidolite would contain only about 0.0003% radiogenic strontium. This is a very small quantity, and hence what little nonradiogenic strontium there is will be present in extremely minute quantities.

All these available mass spectrographic analyses show that in lepidolite radiogenic strontium is likely to predominate completely, and that the proportion of nonradiogenic strontium will in most instances be insignificant. The age determinations given later in this paper are on lepidolite, and with few exceptions the ages are of the correct magnitude, which further corroborates the mass spectrographic evidence.

HYDROTHERMAL PEGMATITIC MICROCLINE: Hydrothermal pegmatitic microcline is much rarer than primary pegmatitic microcline, which is unfortunate because available evidence shows that in the former the proportion of radiogenic strontium is likely to be high, whereas in the commoner primary type the proportion is usually extremely low. This difference is to be expected; hydrothermal microcline, being of a late replacement phase in a pegmatite, is enriched in rubidium and impoverished in magmatic strontium, whereas the opposite will hold for the primary type which is of much earlier crystallization.

Isotope abundance measurements on strontium from three specimens of hydrothermal microcline (Mattauch, 1947) are given.

Source	Proportion of radiogenic strontium
Varuträsk, Sweden (1)	84.3%
Varuträsk, Sweden (2)	79.5%
Utö	46.3%

In each instance, the proportion of radiogenic strontium is very high.

AMAZONITE: Mattauch (1947) has made an isotope analysis of one specimen of amazonite, from Skuleboda, Sweden, and found 75.7% of the strontium to be radiogenic.

Some indirect evidence supports the probability that in amazonite, like hydrothermal pegmatitic microcline the proportion of radiogenic strontium is likely to be high.

Tolmacev and Filippov (1935) determined strontium in two specimens of amazonite from the Ilmen Mountains, Russia, and found 2.0% Rb and 0.002% Sr in one specimen, 1.3% Rb and 0.0015% Sr in the other. They assumed an age of about 200×10^6 years for the age of these specimens and, using the old value of 4.3×10^{11} years as the half life of rubidium, they concluded that in each specimen about 30% of the strontium was radiogenic. If the newer half-life value of 5.8×10^{10} years for Rb^{87} is used, the proportion of radiogenic strontium in each specimen is about 65%.

The strontium 'age' of amazonite from Pike's Peak, Colorado, has been determined as 1100×10^6 years (Ahrens, 1947a). Lead ages on pegmatites of the same age are usually about 1000×10^6 years, whence the proportion of radiogenic strontium is approximately 90%. Other observations on radiogenic strontium in amazonite are given by Ahrens (1947c).

POLLUCITE: A quantitative spectrochemical analysis and a consideration of some aspects of the geochemistry of strontium (Ahrens, 1945b) suggests that the strontium in a specimen of pollucite from Karibib, South West Africa, was largely radiogenic. Hahn *et al.* (1943) had shown mass spectrographically that in a specimen of pollucite from Varuträsk, 76.5% of the strontium was radiogenic, but, because of wartime conditions, the mass spectrographic analysis was not known to me. As a result of further research on more specimens of pollucite from other localities, it was concluded by Ahrens, (1947b) that in these specimens too, the strontium was probably largely radiogenic.

Pollucite is characteristic of late hydrothermal replacement phases in pegmatite, but is unfortunately rare.

LITHIUM-RICH MUSCOVITE: Only on the three minerals just discussed have mass spectrographic analyses been made on the relative abundance of the strontium isotopes. In three specimens of lithium-rich muscovite, indirect evidence has shown a high proportion of radiogenic strontium. As in the case of the Pike's Peak amazonite, strontium 'ages' were made on the three specimens of muscovite, and by compar-

TABLE 1.—*Analyses (Bray, 1942)*

Granite type	% SrO	
	Parent rock	Pegmatite
Boulder Creek granite	0.063	0.0055
Overland Mt. granite	0.0098	0.009
Silver Plume granite	0.015	0.0066

ing these 'ages' with the probable ages, the proportion of radiogenic strontium can be calculated approximately. In one specimen (Usakos, South West Africa) about 90% of the strontium was calculated as radiogenic, and in one from the Game Reserve, eastern Transvaal, South Africa, the proportion was determined as 70%. In a specimen of sulphur-yellow muscovite from Keystone, Black Hills of South Dakota, the proportion of radiogenic strontium was calculated as 60%, on the assumption that the age of the mineral was about 850×10^6 years.

Although these results are hardly representative they do show that the proportion of radiogenic strontium in lithium-rich muscovite is high in many instances. In other muscovite, the proportion of radiogenic strontium is almost certainly smaller; available analyses of strontium in ordinary muscovite frequently show a concentration of strontium in considerable excess of what could have formed from the decay of rubidium (Table 1). Although the rubidium content of these muscovites is not known, calculation will reveal that most of the strontium must be nonradiogenic, although the proportion of radiogenic strontium in the pegmatitic muscovite may be appreciable and could be as high as about 20%. The age of the different granites and pegmatites may be taken as roughly equal to 1000×10^6 years.

Several analyses of strontium in muscovite have indicated that the proportion of radiogenic strontium in pegmatitic muscovite is likely to be higher than in muscovite from the associated parent granite (Table 1).

In each case, the concentration of total strontium decreases in the pegmatitic muscovite. Shimer (1943) has observed similar trends. In 15 pairs of granite and associated pegmatitic mica (muscovite and biotite) from New England, which were analyzed semi-quantitatively, the strontium content decreased in the pegmatitic mica and in four instances the strontium content remained about constant; in no instance did strontium increase in the pegmatitic mica. Nothing is known about the rubidium contents of this 15 pairs; some rubidium will be present and will be likely to be more abundant in the micas from the pegmatites. Even if, however, the rudibium content remains about constant, the decrease of total strontium in the pegmatitic muscovite will mean an increase in the proportion of radiogenic strontium.

RHODIZITE AND CESIUM BERYL: An approximate analysis of strontium in the specimen of rhodizite from Manjak, Madagascar, gave about 0.01% SrO; in the associated beryl, about 0.004% SrO was found. To my knowledge, no definite age determinations have been made on these pegmatites, but they are known to be pre-Cambrian, and calculation would indicate that, in both minerals, 20–80% of the strontium is radiogenic. It may be possible to make age determinations on this type of beryl, which is not very rare.

BIOTITE, AND SOME OBSERVATIONS ON COMMON MICROCLINE: Some pegmatitic biotites contain a relatively high content of rubidium, and it seems likely that unless the amount of ordinary strontium is unusually high, the proportion of radiogenic strontium will be very considerable.

Bray (1942) has determined strontium in several specimens of biotite from granite and associated pegmatite (Boulder Creek, Overland Mountain, and Silver Plume granites) from Colorado. The strontium content varies a lot, but is very low, 0.003 and 0.006%, in two specimens of pegmatitic biotite. Since these rocks are pre-Cambrian and about 1000×10^6 years old, probably an appreciable proportion of the strontium is radiogenic. The total strontium content in most of the granite biotites is in general much higher and the proportion of radiogenic strontium undoubtedly much lower.

Recent analyses (Nockolds and Mitchell, 1948) of strontium and rubidium in biotite in Caledonian granites from Britain are interesting. On the assumption that these rocks are about 350×10^6 years old the proportion of radiogenic strontium may be calculated as 1–5% in most cases. The significance of these results is discussed later.

Where mica (muscovite or biotite) is associated with microcline (not the hydrothermal variety) much more strontium appears to concentrate in the feldspar. From analyses (Bray, 1942) of strontium in mica: feldspar pairs in granite and in pegmatite, the ratio %SrO in mica/%SrO in feldspar is usually 1/10. Qualitative observations (Shimer, 1943) also showed an increase of strontium in the feldspar relative to mica. Calculating from the analytical data of Nockolds and Mitchell (1948), the proportion of radiogenic strontium in potash feldspar from the Caledonian granites averages about 0.01%, and is thus about 100 times lower than in the associated mica. The mean ratio, %SrO in mica/%SrO in feldspar, is about 1/25.

OTHER MINERALS: The proportion of radiogenic strontium in some phlogopite specimens is considerable (Ahrens, 1948a) and it seems reasonable that in some instances this mineral could be used for the determination of geologic age.

To sum up, available data show that in lepidolite, the proportion of radiogenic strontium is likely to predominate completely, and that it should therefore be possible to make reasonably accurate age determinations on this mineral in many instances without a mass spectrographic measurement of the relative abundance of the different strontium isotopes; naturally such an analysis is desirable. In hydrothermal microcline from pegmatite, amazonite, pollucite, some specimens of lithium-rich muscovite, possibly also lithium-rich biotite, and probably rhodizite and cesium beryl, the proportion of radiogenic strontium is likely to be very high, probably within the limits of 15–90%. In other minerals, the proportion of radiogenic strontium

is likely to be much lower, with the possible exception of phlogopite; in common microcline, the proportion is very low.

A more complete and detailed discussion of the geochemistry of radiogenic strontium is given by Ahrens (1948a).

QUANTITATIVE DETERMINATION OF Sr/Rb RATIOS

GENERAL

The usefulness of the strontium method will depend, naturally, to a very large extent on the ease and accuracy with which quantities of rubidium from 2.5—about 0.05% and even less, and of strontium from 0.02–0.0002%, can be determined.

Chemical analysis has apparently been successful for estimating rubidium in lepidolites and in other minerals where the concentration of rubidium is less (Stevens and Schaller, 1942), but amounts of about 0.1% and less would be difficult to determine chemically. Amounts of strontium less than 0.02% have been determined chemically by Strassmann (Strassmann and Walling 1938; Hahn *et al.*, 1943), but the procedure is somewhat lengthy, and considerable difficulty would be experienced in determining quantities of strontium as low as 0.001%, which amount is average for lepidolite of an intermediate age, namely about Ordovician. In addition, a relatively large bulk of the mineral is required for the chemical analysis and some significant loss of strontium is likely during chemical manipulations. Thus, in the ancient lepidolite from Manitoba, Canada, which is the most ancient thus far analyzed and which contains about 0.015% Sr, the recovery of strontium may be 10% too low (Strassmann and Walling, 1938). In younger minerals and in minerals which contain smaller amounts of rubidium, greater loss of strontium is to be expected.

Both rubidium and strontium may be determined spectrochemically. Rubidium has been determined by several different techniques (Goldschmidt, *et al.*, 1935; Tolmacev and Filippov, 1935; Rusanov and Vasil'ev, 1939; Adamson 1942; Ahrens, 1945a; and others) and no serious difficulties have been reported. Using some refinements in technique, it should be possible to determine this element with a reproducibility of ±5.0%.

The determination of strontium in minerals and rocks has been, in most instances, by semi-quantitative methods. Noll (1934) describes a method he developed for investigating the geochemistry of strontium.

Strock (1942) carried out investigations on the determination of strontium in some specimens of lepidolite analyzed by Stevens (1938). Using a series of standards prepared from a base identical in composition to that of the mineral, excellent reproducibility is reported, using Sr 4607 with a lithium line as an internal standard line. As lithium is used as an internal standard and because the lithium content of lepidolite varies markedly in different specimens, one would have to know the lithium content beforehand in order to use this method.

More recently, Hybbinette (1943) combined a preliminary chemical concentration of strontium (precipitation of strontium as oxalate in the presence of a large excess of calcium) followed by a later spectrochemical analysis of the strontium in the ignited precipitate, using calcium as an internal standard. The method is interesting

and the accuracy reported good, but a serious loss of strontium may be expected when the strontium content of the mineral is low. Hybbinette's statement that the chemical enrichment is required because a spectrochemical analysis cannot be made directly when the concentration of strontium is below 0.01% requires some comment, because strontium is a sensitive element spectrochemically and less than a tenth of this amount can invariably be detected and estimated.

DEVELOPMENT OF A SPECTROCHEMICAL PROCEDURE

Purpose.—The author has attempted to determine rubidium and strontium in a single operation. In this method to be described, an attempt is made to use self-internal standardization, that is, the ratio Sr/Rb is determined from the intensity ratio of a pair of strontium and rubidium lines.

Apparatus.—Two spectrographs have been at my disposal; a larger Hilger quartz spectrograph of the Government Metallurgical Laboratory, University of the Witwatersrang, Johannesburg, and a large Wadsworth type grating spectrograph (linear dispersion 2.5 A/mm) of the Cabot Spectrographic Laboratory, Department of Geology, Massachusetts Institute of Technology, Cambridge.

The material to be analyzed was placed in the anode, of purest available graphite or carbon, the dimensions of which were; external diameter $\frac{3}{16}''$, internal diameter $\frac{1}{8}''$ and depth $\frac{1}{4}''$. The cathode, of similar material, was $\frac{3}{16}''$ in diameter and was pointed.

A d.c. arc source (line voltage 220 volts) was used, and after a pre-arcing period of about 2 seconds at low amperage, the current was increased to 6–7 amps, and the specimen arced to completion. The spectra were recorded on Ilford Ordinary plates and also on Eastman Spectroscopic plates, Type 103-0, both of which were found suitable, and were developed under standardized conditions for 4 minutes at 18°C in Kodak D 19 developer. After fixing, washing, and drying, microphotometric measurements were made on Hilger nonrecording microphotometers, one at the Government Metallurgical Laboratory and one at the Dept. of Physics, Massachusetts Institute of Technology.

Only the purest electrodes were used (Hilger's and National Carbon) because more impure grades revealed the presence of strontium lines of considerable intensity.

A stepped sector of seven steps having a transmission factor of two was used to obtain line intensity ratios and to decrease the intensity of Rb 4202, which in lepidolite is very intense.

Selection of suitable analysis lines.—The most sensitive lines of rubidium are located in the red end of the spectrum, but these lines were disregarded for this work because at relatively large concentrations of rubidium; these ground state lines are prone to reverse excessively; furthermore, their location is not convenient. Rb 4202 and Rb 4215, less sensitive lines, can be used to determine concentrations of rubidium as low as 0.005%, which sensitivity is ample; but Rb 4215 could not be used because it is co-incident with Sr 4215 and the head of a CN band. Rb 4202 was therefore selected provisionally, and wave-length tables were examined for the presence of interfering lines (Table 2). Only lines listed as having an intensity greater than "50" are given.

Only Fe 4202 has been found to cause interference; the weaker iron line, Fe 4200.97, has not been observed. Experience has shown that interference from Fe 4202 is very rare in lepidolite, absent in feldspar, but common in muscovite and biotite. If the dispersion of the spectrograph is large enough (~ 3.0 A/mm) and if the amount of iron is not excessive, as in biotite but not in muscovite, Fe 4202.03 and Rb 4201.85

TABLE 2.—*Wave-lengths and intensities of lines neighbouring Rb 4202 (Harrison, 1939)*

Element	Wave length	Intensity
Tm	4203.73	250
Ti	4203.46	50
Os	4202.06	100
Fe	4202.03	400
Rb	4201.85	2000
Zr	4201.46	50
Fe	4200.97	80
Cr	4200.13	100
Ru	4199.90	150

should appear clearly resolved on the plate and the amount of interference is then usually negligible. If the dispersion is small, as for example in quartz spectrographs, the two lines nearly coincide and correction can be made as described later.

Strontium has three sensitive lines situated reasonably close to Rb 4202; Sr 4077.71, Sr 4215.52, and Sr 4607.33. Of these lines Sr 4607 is the most sensitive but is located close to an intense lithium line, Li 4603, and, since lepidolites contain on an average 4–5% Li_2O, this line is extremely intense when lepidolite is arced. Consequently, unless the dispersion is very large, that is, considerably greater than in large quartz spectrographs, interference from Li 4603 is excessive, particularly in young lepidolites where the strontium content varies from about 0.0002–0.001%. If the dispersion is about 2.5 A/mm, which is the dispersion of the Cabot Spectrographic laboratory instrument, interference is slight except in young lepidolites, but correction can usually be made, whereas if instruments of very large dispersion are used, for example the 3-meter Paschen 30,000 line per inch grating spectrograph of the Massachusetts Institute of Technology physics department, no interference at all is encountered. The linear dispersion of this instrument is about 0.85 A/mm.

Since only quartz optics were available at the commencement of this research, Sr 4607 could not be used; Sr 4215 was disregarded because it is roughly coincident with Rb 4215. Only Sr 4077 was left to be investigated.

Only Mn 4079 (doublet) and Cb 4079 have been detected in lepidolites so far examined. Reference to photographs (Ahrens, 1947a) show that even in quartz optics, these lines, although closely situated on the plate, do not cause actual interference.

Ti 4078 has not been detected in lepidolite, but has been detected in some specimens of muscovite. Using quartz optics, Ti 4078 interferes seriously with Sr 4077,

but correction may be made. When using the grating spectrograph, no significant interference has been observed.

Both Rb 4202 and Sr 4077 are within a cyanogen band, the head of which is at 4216 A, and since this band is an intense one, cyanogen emission has to be effectively

TABLE 3.—*Wave lengths and intensities of lines neighboring Sr 4077 (Harrison, 1939)*

Element	Wave length	Intensity
Cb	4079.73	500
Mn	4079.42	50
Mn	4079.24	50
Ti	4078.47	125
Fe	4078.36	80
Dy	4077.97	150
Hg	4077.81	150
Sr	4077.71	400
Co	4077.41	100
Y	4077.37	50
La	4077.34	600
Ru	4076.73	60
Fe	4076.64	80
Co	4076.13	70

quenched lest it interfere excessively. It will be shown later that in the presence of much alkali metal vapor, CN emission can be reduced to a negligible minimum and satisfactory measurements can be made even on faint lines.

Method of excitation.—Direct current arc excitation is frequently used for the analysis of powders; some operators prefer to place the specimen in the anode (anode excitation) whereas others employ the cathode, and some record only that portion of the arc immediately adjoining the cathode (cathode layer excitation). Selection of method is generally conditioned by the particular element(s) sought, the type of material to be analyzed, and the quantity available for analysis.

Since Rb 4202 and Sr 4077 are located in a cyanogen band region and since under favorable conditions anode excitation produces only a bare minimum of cyanogen emission, the use of this method was invesitgated.

Because the author attempts to employ the principle of internal standardization, it would be well to review here the criteria governing the selection of a good internal standard. The Criteria, in this instance, are:

(1) The lines to be used should if possible be situated reasonably close to each other,

(2) Sr 4077 and Rb 4202 should have similar potentials of excitation, and

(3) Strontium and rubidium should volatilize from the specimen in a like manner.

Condition (1) is satisfied, but condition (2) is not, since Sr 4077 is a spark line with an excitation potential of 8.7 volts, and Rb 4202 is an arc line with an excitation of only 2.9 volts. The absence of experimental data makes it difficult to assess the

influence of this difference on the reproducibility of Sr 4077/Rb 4202 intensity ratios. As the difference in excitation potential is great, the consequence will undoubtedly be serious, but more will be said about this later.

When direct current arc anode excitation is employed, various elements volatilize into the arc column with very marked differences, particularly in silicates where much alkali metal is present; therefore, this factor of volatilization (condition (3)) is probably the most important one in determining the reproducibility of Sr 4077/ Rb 4202 intensity ratios.

A specimen of lepidolite was arced to completion and the plate holder of the spectrograph racked down at intervals of 15 seconds. From Figure 3 it is clear that rubidium, (to-gether with the other alkali metals the volatilization curves of which are not shown), is relatively volatile, whereas the appearance of Sr 4077 is very late, and it is obvious that Sr 4077 and Rb 4202 do not provide each other any internal standardization.

Apart from the factor of internal standardization, the late appearance of the strontium spectrum is unsatisfactory for two reasons. First, cyanogen emission is at a minimum during the early period of arcing when an abundance of alkali metal vapour is present in the arc column, but after the bulk of the alkali metals have volatilized off, cyanogen emission becomes much more intense and Sr 4077 emits its peak intensity of emission (Fig. 3). Cyanogen emission is particularly intense during these final stages of arcing because the residual bead in the anode consists predominantly of alumina which is not readily arced and the arc tends to 'play' around the alumina bead consuming large amounts of carbon instead of striking the bead. Second, this residual bead which contains the bulk of the strontium may not always be arced to completion successfully because it has the very annoying tendency of being ejected out of the anode cavity.

Consequently, unless strontium could be made much more volatile and cyanogen emission drastically decreased, there is scant hope of determining Sr/Rb ratios from measurements of Sr 4077/Rb 4202 intensity ratios. Admixture of powdered carbon with the material to be arced partly eliminates selective volatilization, and the use of a carbon anode rather than graphite aids a little. Cyanogen emission is still excessive, however, particularly for young lepidolites.

In silicate minerals it is common to regard the volatilities of the elements as that of their oxides. If the elements are converted into volatile compounds, for example the halides, all elements are made relatively volatile. A conversion to fluorides was therefore attempted. A specimen of lepidolite was treated with hydrofluoric acid, taken to dryness twice carefully, and arced to completion as before. The results were markedly different (Fig. 4): all elements were rendered relatively volatile and cyanogen emission was negligible because a high concentration of alkali metal vapour was available throughout the whole period of arcing. Although there still is a considerable difference in volatilization rates, the difference is very much smaller than without the addition of hydrofluoric acid and the trends of the two curves are sufficiently close to warrant the use of internal standardization. Unfortunately the individual intensity fluctuations across the curve do not appear to be sympathetic.

To ascertain the reproducibility of the Sr 4077/Rb 4202 intensity ratios, a speci-

men of lepidolite was arced 16 times, after treatment with hydrofluoric acid and taking to dryness. The results (Fig. 5) are rather disappointing because of the spread of points, and the approximate standard deviation from the mean is about ± 30–40%. If each analysis is carried out in quadruplicate, a standard deviation in the repro-

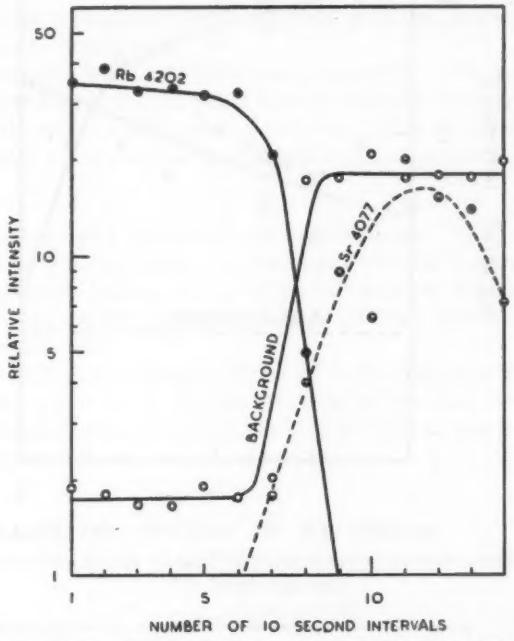


FIGURE 3.—*Selective volatilization curves for Sr 4077 and Rb 4202 in lepidolite*

Obtained by plots of line intensity vs. number of fifteen second intervals.

ducibility of about ± 10–15% may be obtained, which is satisfactory for preliminary investigations and many other investigations which do not require very precise measurements. Consequently it was decided to attempt the measurement of Sr/Rb ratios by measuring Sr 4077/Rb 4202 intensity ratios after each specimen had been converted to fluorides.

Preparation of a synthetic standard.—For accurate quantitative analysis of powders, the use of synthetic standards is avoided if possible, because the intensity ratio of a given line pair at a given concentration of an element in a synthetic standard is likely to differ from the intensity ratio of the same line pair at the same concentration of the element in the natural mineral. This error is omnipresent in spectrochemical analysis to some degree, however precise the method, and the magnitude of error is often difficult to assess. In this instance, synthetic standards had to be prepared, but as lepidolite is broken down into a mixture of fluorides on treatment with hydrofluoric acid, the similarity between the synthetic standard and that of the acid-

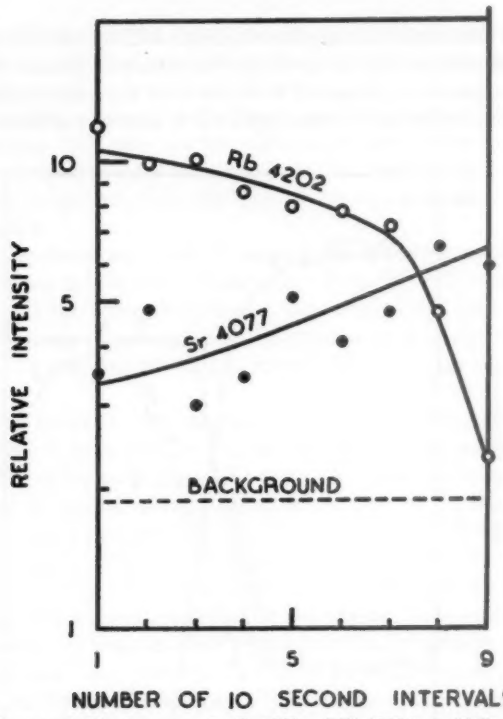


FIGURE 4.—*Selective volatilization curves for Sr 4077 and Rb 4202 in lepidolite after treatment with hydrofluoric acid*

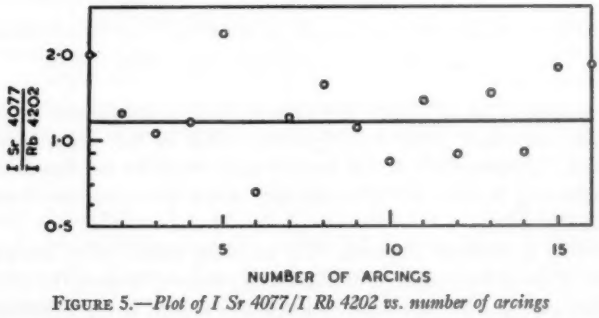


FIGURE 5.—*Plot of I Sr 4077/I Rb 4202 vs. number of arcings*

The straight line drawn through the points gives the mean. Points are widely scattered about the mean.

treated mineral is undoubtedly very close if the synthetic standard is prepared from a mixture of fluorides.

The ratio of strontium to rubidium was adjusted to be equivalent to about 1400 $\times 10^6$ years in a synthetic standard of the following composition:

AlF_3	57.6%
LiF	15.0%
KCl	25.0%
RbCl	2.4%
SrF_2	0.011%

A single standard, corresponding to a relatively great age, was used for reasons which are discussed a little later.

Relationship between I Sr 4077/I Rb 4202 and geological age.—The basic assumption in quantitative spectrochemical analysis is that the intensity (I) of a given line of an element is directly proportional to the concentration (C) of the element within the source of excitation. The relationship is usually expressed as follows:

$$I = KC^n \quad (5)$$

where K is a constant and n is known as the emission factor. The value of n is frequently close to unity and is equal to it when there is no self-absorption of energy: n is dependent on several factors, such as the type of source and the energy conditions prevailing within the source, the concentration of emitting atoms, and the energy characteristics of the line.

Let I_{Sr} = I Sr 4077, I_{Rb} = Rb 4202, n′ and n″ = the respective emission factors of these two lines, and let C_{Sr} = the concentration of strontium within the source and C_{Rb} = the concentration of rubidium: let k′ and k″ refer to constants for Sr 4077 and Rb 4202, respectively.

Then,

$$I_{Sr} = k' \cdot C_{Sr}^{n}$$

and

$$I_{Rb} = k' \cdot C_{Rb}^{n''},$$

whence

$$I_{Sr}/I_{Rb} = k'/k'' \cdot C_{Sr}^{n'}/C_{Rb}^{n''} \quad (6)$$

Replace k′/k″ by K, and introduce logarithms,

$$\log \frac{I_{Sr}}{I_{Rb}} = \log K + \frac{n'}{n''} \log \frac{C_{Sr}}{C_{Rb}} \quad (7).$$

In a plot of log I_{Sr}/I_{Rb} vs. log C_{Sr}/C_{Rb}, the ratio n′/n″ will determine the slope of the graph, and if this ratio is equal to unity, the slope of the graph will be unity.

As a large blank error for strontium was found for synthetic standards corresponding to relatively young minerals and could not be corrected for accuracy, and because the strontium blank error was very slight in standards corresponding to old minerals and correction could be made easily, only this single synthetic standard was used by means of which a working curve (Fig. 6) of unit slope was prepared: the preparation of this curve assumes a value of unity for the ratio n′/n″.

There is very little likelihood that n′ will deviate from unity by more than a negligible amount over the range of concentrations investigated (0.0003–0.02% Sr) because Sr 4077 is a high energy (spark) line and the quantities of strontium in the above range are very small. Unlike Sr 4077. Rb 4202 is a low-energy line and ru-

bidium is present in much larger quantities; consequently some self-absorption is probable. Personal observations on this line have shown that the amount cannot be very great, however. Evidence of the high reproducibility of Rb/K ratios in lepidolite (referred to later) based on K 4047/Rb 4202 intensity measurements, shows that self reversal in Rb 4202 cannot be very marked, since the reproducibility would have been much poorer because of the random and uncontrollable self-reversal changes in a line showing strong self reversal. Also, the concentration of rubidium in lepidolite varies over narrow ranges only and the rubidium content of the standard was made approximately equal to that commonly found in lepidolite. In minerals where the rubidium content is relatively low, a small deviation of n″ from unity would have serious consequences, and the synthetic standard has only been used for lepidolite and a few other minerals relatively rich in rubidium.

Assuming n′ and n″ each to equal unity, equation (6) becomes,

$$\frac{I_{Sr}}{I_{Rb}} = K \times \frac{C_{Sr}}{C_{Rb}} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots \quad (8)$$

Combining equations (3) and (8) gives,

$$\text{Age} = \frac{\frac{I_{Sr}}{I_{Rb}} \times 3.33 \times 10^{12}}{K} \text{ years} \quad \ldots\ldots\ldots\ldots\ldots\ldots \quad (9)$$

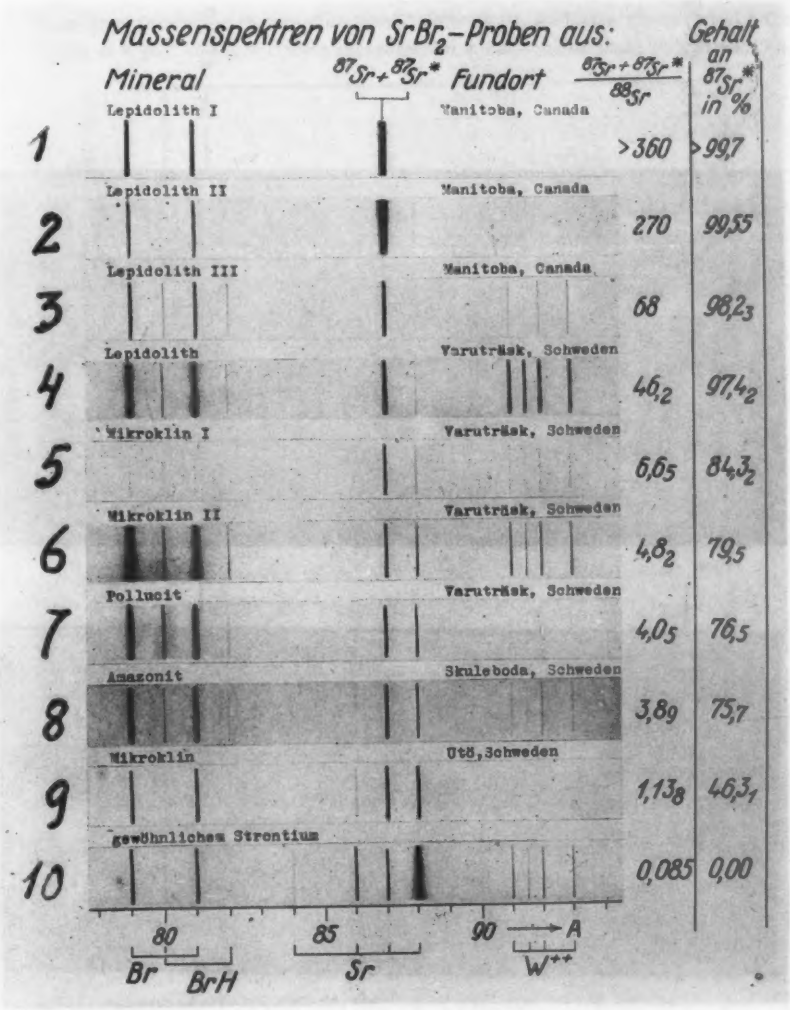
or

$$\log \text{Age (years)} = \text{Log} \frac{I_{Sr}}{I_{Rb}} + \log (3.33 \times 10^{12}) - K \quad \ldots\ldots\ldots\ldots \quad (10)$$

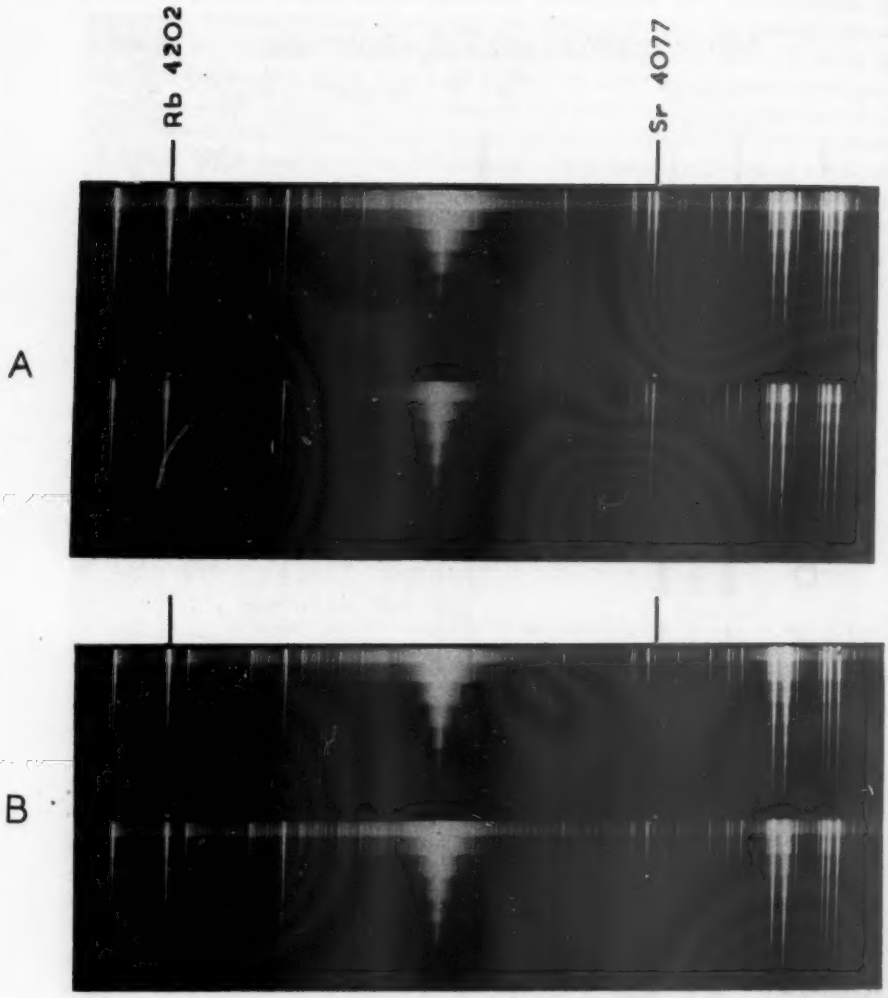
Equation (10) shows that it is not necessary to determine the actual quantities of rubidium and strontium, since the intensity ratio Sr 4077/Rb 4202 is directly proportional to the age of the mineral.

The ratio I Sr 4077/I Rb 4202 was determined accurately (mean of eight determinations) for the synthetic standard, the Sr/Rb ratio of which was equivalent to 1400×10^6 years, and a plot was made of the line–intensity ratio vs. the age: a straight line of unit slope through this point (Fig. 6) furnished the working curve to which I Sr 4077/I Rb 4202 ratios of the unknowns are referred for an estimation of their ages.

Microphotometric measurement of the ratio I Sr 4077/I Rb 4202.—Since a stepped sector was rotated in front of the slit of the spectrograph, each line recorded on the plate shows the presence of several steps of varying density (Pl. 2). From these stepped spectrograms, I Sr 4077/I Rb 4202 ratios were determined as follows: The densities of two or more steps of Sr 4077 and Rb 4202 are determined and plotted as ordinates against each corresponding sector step. Steps were chosen for each line which provided density values above and below 1.0. Density values of less than 0.4–0.5 were not plotted because they fall on the toe of the characteristic curve and not on the straight line portion. In the resultant curves, which are partial characteristic curves, (Fig. 7), the step number of the step sector is made to increase from left to right, where step number one is the step of minimum transmission; consequently,


Massenspektren von $SrBr_2$-Proben aus:
Mineral
$^{87}Sr + ^{87}Sr^*$
Fundort
$\frac{^{87}Sr + ^{87}Sr^*}{^{88}Sr}$
Gehalt an $^{87}Sr^*$ in %
1 Lepidolith I Manitoba, Canada >360 >99,7
2 Lepidolith II Manitoba, Canada 270 99,55
3 Lepidolith III Manitoba, Canada 68 98,2$_3$
4 Lepidolith Varuträsk, Schweden 46,$_2$ 97,4$_2$
5 Mikroklin I Varuträsk, Schweden 6,65 84,3$_2$
6 Mikroklin II Varuträsk, Schweden 4,8$_2$ 79,5
7 Pollucit Varuträsk, Schweden 4,05 76,5
8 Amazonit Skuleboda, Schweden 3,89 75,7
9 Mikroklin Utö, Schweden 1,13$_8$ 46,3$_1$
10 gewöhnliches Strontium 0,085 0,00
80 85 90 A
Br BrH Sr W^{++}

MASS SPECTRA (MATTAUCH, 1947) OF STRONTIUM IN SEVERAL RUBIDIUM-RICH MINERALS



DUPLICATE SPECTRA OF HYDROFLUORIC ACID RESIDUES OF A, AN ANCIENT LEPIDOLITE OF 2100 X 10^6 YEARS, AND B, ONE OF 100 X 10^6 YEARS

lines producing partial characteristic curves displaced to the left are of greater intensity than lines producing curves displaced to the right, and at a given density the separation of a pair of these curves measured on a convenient scale determines the

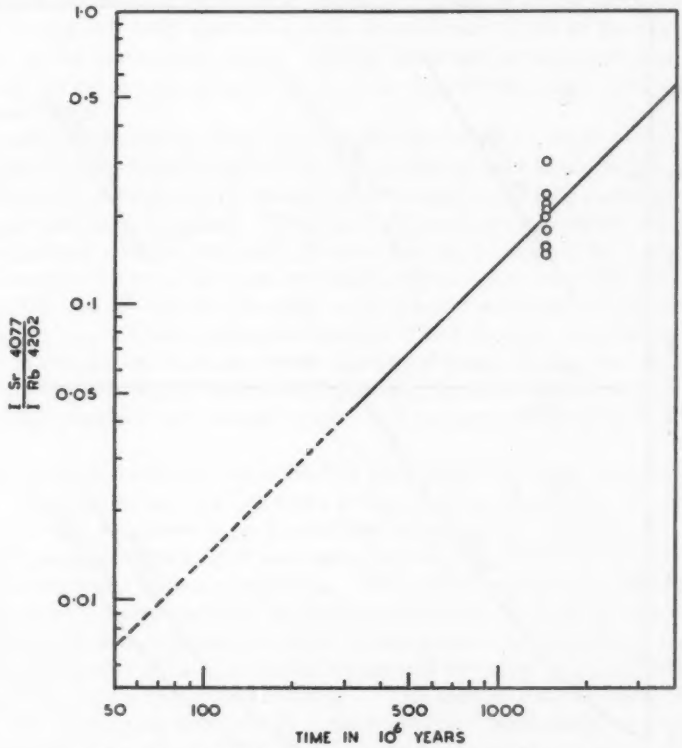


FIGURE 6.—*Working curve: plot of I Sr 4077/I Rb 4202 vs. age*

This curve of unit slope assumes a value of unity for n'/n''.

intensity ratio of the two lines. This method is commonly used, but a slight modification introduced in this paper is useful in making correction for background. An arbitrary scale of relative exposures (intensities) is used which is reversed relative to the stepped sector apertures, that is, the largest exposure values are on the left (Fig. 7). At a given density (a density of 1.0 was chosen) on the straight line portion of the characteristic curve, relative exposure values corresponding to each line are read off on the relative exposure scale. The intensity ratio of a pair of lines is then the ratio of the exposure values read off on the absicissa.

The photographic response, measured as density here, of a line lying in a background is produced by the intensity of the line plus that of the background emission. Pierce and Nachtrieb (1941) have shown that a subtraction of background intensity from the total intensity, that is of line plus background, corrects accurately for the

influence of background, and personal experience has corroborated their conclusions in many instances. The procedure adopted here is as follows: On suitable steps, density measurements are made of the background immediately adjacent to the line;

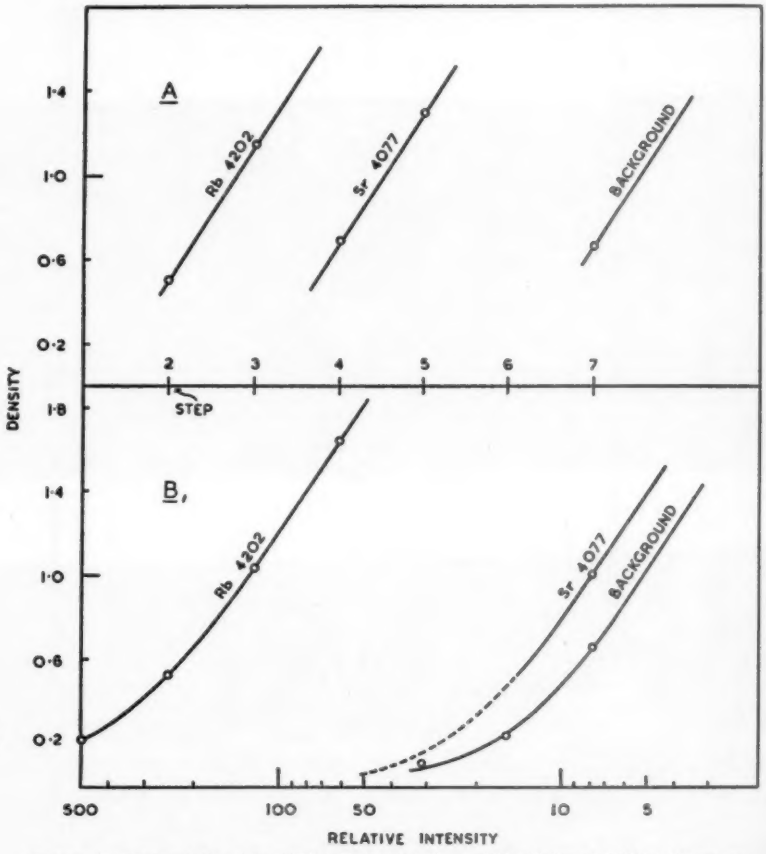


FIGURE 7.—*Determining I Sr 4077/I Rb 4202: (A) an old lepidolite of 2000 × 10⁶ years; (B) a young lepidolite of 100 × 10⁶ years*

plots are made as before and the relative intensity for background is read off on the abscissa at a density of 1.0; the corrected line intensity is obtained by subtracting the background intensity from the total intensity. (Care should be exercised not to subtract the log intensity values.)

For the determination of I Rb 4202 in lepidolite, background intensity was found to be always negligible. For Sr 4077, background was likewise insignificant for very ancient specimens, but serious for young lepidolites; in some instances background intensity was equal to that of the line and consequently accurate correction had to

be made. This was particularly true of quartz optics which were used for the earlier investigations, but when a greater dispersion (2.5 A/mm) of the grating spectrograph was used, background was considerably less in evidence.

In some instances the background density of the step of largest transmission fell below a value of 0.4–0.5, that is below the lowest density to fall on the straight line portion of the characteristic curve. Extrapolation was necessary and was carried out with the aid of a completed pattern of the characteristic curve which had been prepared before.

Correction for interfering lines.—(a) Fe 4202.031 with Rb 4201.85. If the interfering line is of equal wavelength to the analysis line the total photographic response is produced by the intensity of the analysis line plus the intensity of the interfering line plus background intensity. Therefore, if the intensity of the interfering line and the background intensity are known, a correction can be made if the analysis line and interfering line are of the same wavelength or very nearly so. With reference to Fe 4202.031 and Rb 4201.85, in quartz optics where the dispersion is small in this spectral region, the above assumption should not lead to serious error unless the intensity of the analysis line is very weak and that of the interfering line very strong. If the dispersion is greater than about 3.0 A/mm, the two lines are resolved on the plate provided neither is excessively intense, and no correction has been found necessary.

When using quartz optics, correction was made where necessary—usually in muscovite. A neighbouring iron line which is slightly more intense than Fe 4202 was sought; Fe 4143 was found to be of convenient intensity, and employing the normal operating conditions described in this paper, the ratio I Fe 4202/I Fe 4143 was determined as 0.85 in the absence of rubidium. If Rb 4143 is absent in the spectrum of a mineral which is being analyzed, no interference from Fe 4202 will be encountered. If Fe 4143 is present, however, correction for interference may be made by subtracting 0.85 $\times$ intensity of Fe 4143 (minus background intensity) from the intensity of Rb(Fe) 4202 (minus background intensity). This method has been used (Ahrens, 1945a) for determining rubidium in various minerals, particularly in muscovite.

In lepidolite, the presence of Fe 4143 was observed only once and since the spectrum had been recorded using the large dispersion produced by the grating, no interference was encountered. In muscovite, however, the presence of Fe 4143 was always observed, and if quartz optics were used, the required correction was introduced.

(b) Ti 4078.47 and Sr 4077.71. Interference from Ti 4078 was not encountered in any specimens of lepidolite analyzed, but was definitely noticeable in a specimen of muscovite from Usakos, South West Africa. The interference could be ascertained by a broadening of Sr 4077 (using quartz optics), which line was at the same time excessively intense compared with Sr 4607; in addition, other titanium lines were clearly in evidence. Because of a considerable difference in wavelength between Sr 4077.71 and Ti 4078.47, correction could not be made as before when using quartz optics; no interference has been encountered when using the grating spectrograph.

Since only relatively small quantities of lithium are usual in muscovite (0.072% Li_2O in the specimen from Usakos), masking of Sr 4607 by Li 4603 does not occur as in lepidolite and the relative intensity of Sr 4607 may be accurately measured;

whence, if the ratio *I* Sr 4607/ *I* Sr 4077 is known, the relative intensity of Sr 4077 may be calculated. The above intensity ratio was determined in quadruplicate as 4.7, using normal operating conditions.

The following calculations will serve to illustrate an example of how corrections for interference from Ti 4078 and also from Fe 4202 have been made. All intensity measurements were made as described before.

Example of the determination of I Sr 4077/I Rb 4202 in the presence of interference from Fe 4202 and Ti 4078

Determination of I Rb 4202		Determination of I Sr 4077	
I Rb(Fe) 4202 + background	= 5.7	I Sr 4607 + background	= 3.0
I background	= 0.7	I background	= 2.1
Hence, I Rb(Fe) 4202	= 5.0	Hence, I Sr. 4607	= 0.9
I Fe 4143 + background	= 5.0	I Sr 4077 = 1/4.7 × 0.9	= 0.19
I background	= 0.7		
Hence, I Fe 4143	= 4.3		
But I Fe 4202 = 0.85 × Fe 4143	= 3.65		
Whence I Rb 4202	= 1.35		

I Sr 4077/I Rb 4202 = 0.19/1.35 = 0.14. Reference of this intensity ratio value to Figure 6 gives an age of 1000×10^6 years.

Reproducibility of age determinations.—As mentioned earlier, the standard deviation is about 30–40% per single determination; by making each determination in quadruplicate, a 10–15% deviation may be obtained. Agreement between several determinations is very close in some instances and poor in others. When agreement is very poor, the specimen is usually re-analyzed. No reason has been found for the apparent greater reproducibility for some specimens; the degree of homogeneity is a possible cause. The different Sr 4077/Rb 4202 intensity ratios given in Table 6 will give the reader some indications of the reproducibility of the age determinations.

Some further investigations on the quantitative determination of Sr/Rb ratios.—All the age determinations made by the author and given in this paper have been determined according to the method already outlined; subsequent to the determination of these ages, further investigations were carried out in an endeavor to increase the accuracy with which Sr/Rb ratios could be measured.

It will be recalled that Sr 4077 is a spark line and Sr 4607, an arc line, could not be used because of the excessive interference from Li 4603 when quartz optics were used. Using the larger dispersion of the grating spectrograph, however, observations could be made on Sr 4607.

A hydrofluoric acid residue of lepidolite was arced to completion and the plate holder racked down at intervals of 20 seconds (Table 4).

The intensity measurements on all three lines in runs No. 3, 4, and 5 (Table 4) (in the other instances, Sr 4607 was sometimes excessively intense) show:

(1) The intensity of Sr 4607 seems to vary more sympathetically with Rb 4202 than does the intensity of Sr 4077,

(2) Although Sr 4077 and Sr 4607 are lines of the same element and therefore no volatilization factor will influence the intensity ratio of one to the other, the intensity variations of these two lines are not very sympathetic; this is shown clearly in Table 5

which gives the ratio I Sr 4607/I Sr 4077 during the first interval and during the last: the ratio (mean of all values) changes by a factor of 6.8.

(3) In one instance (No. 4 in Table 4), both Sr 4077 and Sr 4607 become very much

TABLE 4.—*Intensity values*
For Sr 4607, Sr 4077, and Rb 4202 during 20 second intervals

	No. of 20 second intervals	Intensity		
		Sr 4607	Sr 4077	Rb 4202
1	1	29.5	13.8	27.0
	2	d	60.0	60.0
	3	d	32.3	31.6
	4	41.8	52.5	7.4
	5	—	—	—
2	1	14.0	4.0	7.6
	2	d	83.0	83.0
	3	d	150.0	19.0
	4	41.8	69.0	15.0
	5	9.5	27.0	—
3	1	38.0	15.0	18.7
	2	250.0	48.0	83.0
	3	74.0	84.0	28.0
	4	28.0	31.6	10.0
	5	5.5	16.0	—
4	1	15.5	3.5	5.0
	2	160.0	28.0	125.0
	3	18.4	31.0	28.0
	4	35.6	85.0	—
	5	—	—	—
5	1	21.0	4.5	16.0
	2	30.3	5.9	18.7
	3	52.4	38.0	24.0
	4	45.0	38.0	16.4
	5	13.0	21.0	—

"—" indicates a line to be absent; "d" that the line is excessively black for microphotometric measurement.

more intense than Rb 4202 during the last period of arcing; this will be discussed in detail later.

The relatively sympathetic behavior of Sr 4607 and Rb 4202 in Nos. 3 and 5 raised hopes that a more reproducible determination of Sr/Rb ratios could be made by using Sr 4607, because, being an arc line having an excitation potential (2.7 volts) very nearly the same as that of Rb 4202 (2.9 volts), a variation of energy conditions within the source should cause a like response from these two lines.

Since spectra of many of the specimens of lepidolite had been recorded using quartz optics and hence Sr 4607 could not be used directly, I attempted to make corrections for unsympathetic intensity variations of Sr 4077 relative to Rb 4202 by making

observations on the intensity variations of two calcium lines, Ca 4226, an arc line (excitation potential 2.9 volts) and Ca 3933, a spark, line (excitation potential 9.2 volts)—it was hoped that the two calcium lines could be used as a control pair. Investigations by Rollwagen (1939) had shown that the intensities of Ca 3933 and Sr 4077 (the two spark lines) vary relatively sympathetically in the arc, as do the intensities of the pair Ca 4226 and Sr 4607 (the two arc lines) when the energy conditions of excitation change within the source; a high temperature favors higher intensities for Ca 3933 and Sr 4077 relative to Ca 4226 and Sr 4607. These sympathetic variations of intensity have also been observed when the hydrofluoric acid residues of lepidolite have been arced and are very marked.

TABLE 5.—*Changes in ratio I Sr 4607/I Sr 4077*

	Ratio I Sr 4607/I Sr 4077	
	Beginning	End
1	2.3	0.80
2	3.5	0.33
3	2.5	0.36
4	4.6	0.42
5	4.7	0.62
Mean.....	3.4	0.51

Some measurements on the intensity variations of Ca 4226 and Ca 3933 with time in hydrofluoric acid residues of lepidolite showed that this pair of lines behaved very similarly to the pair Sr 4077 and Sr 4607. The mean intensity ratio Sr 4607/Sr 4077 changes by a factor of 6.8 during the intervals of arcing observed (Table 5), and measurements on the intensity ratio changes of Ca 4226/Ca 3933 showed a change of 5.6 (mean of two values). The intensity measurements on the calcium lines could not be made on the same spectra which were used for the strontium line measurements because the lines were too intense for exact comparison. The trends are the same in each case, however, and the magnitude in the change of variation is very similar. Consequently, if a mean value is obtained from a sufficient number of I Ca 4226/I Ca 3933 ratios for lepidolite, deviation of the value of the intensity ratio from the mean value would indicate the necessity for making a correction for the observed value of Sr 4077—an enhancement of Ca 3933 relative to Ca 4226, that is, a decrease in the ratio compared with the mean, should indicate that Sr 4077 is abnormally intense relative to Sr 4607 and hence to Rb 4202, and the amount of correction required should be about equal to the amount of deviation. This method does not take into account the different ways in which strontium and rubidium volatilize, but attempts only to compensate for unsympathetic line responses due to varying energy conditions of excitation within the source.

Thirty four I Ca 4226/I Ca 3933 ratios were made on different specimens of lepidolite (Fig. 8) and the mean ratio was determined as 3.8.

When corrections were applied to I Sr 4077 ratios by multiplying each intensity

value by the calcium line intensity ratio/3.8, only a slight improvement in reproducibility was obtained.

Some other factor was sought to account for the nonreproducibility of the I Sr 4077/I Rb 4202 intensity ratios. Table 4, No. 4 showed that both Sr 4077 and Sr

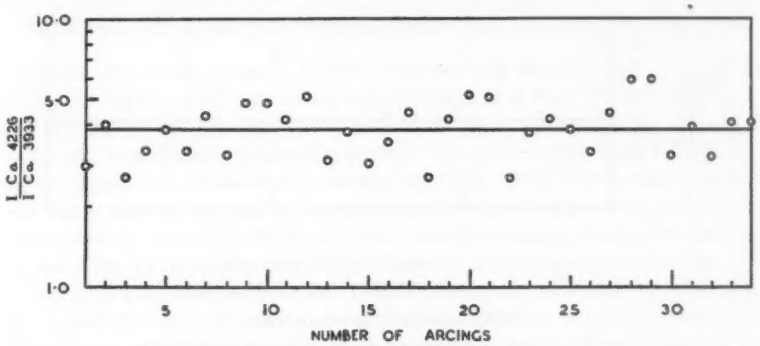


FIGURE 8.—*Plot of I Ca 4226/I Ca 3933 vs. number of arcings*

The spread of points about the mean is very marked.

4607 are very intense relative to Rb 4202 during the last arcing interval, which indicates that strontium does not volatilize smoothly with rubidium in all cases. Further investigation showed that serious differences in volatilization rates between strontium and rubidium were more common than indicated by Nos. 3, 4 and 5 in Table 4, and consequently the use of Sr 4607 or corrected Sr 4077 intensities does not hold much advantage over the use of uncorrected intensity values. The probable deviation of I Sr 4607/I Rb 4202 was determined as about the same as that obtained when I Sr 4077/I Rb 4202 ratios are employed.

Figure 9 shows plots of I Sr 4077/I Sr 4607 for the same analyses given in Figure 5. As in Figure 8 which shows plots of I Ca 4226/I Ca 3933 variations, the distribution of points is very erratic. The standard deviation is about ± 20–30%, which provides a good indication of how serious an error can arise from the use in quantitative analysis of a line pair whose components have very different excitation potentials: since the above intensity ratios refer to a single element in each case, errors due to segregation and differences in volatilization behaviour do not exist, and the deviation is due only to the differences in line response of lines having different energies of excitation.

The statement has sometimes been made that when using d.c. arc anode excitation, the proper reproducibility necessary for accurate quantitative analysis cannot be attained, and this lack of precision has been regarded as one of the most serious drawbacks of using this method of excitation. This view is only correct in part, since good quantitative reproducibility of about $\pm 5\%$ can certainly be obtained under relatively good circumstances. Using d.c. arc anode excitation, the choice of an appropriate internal standard is extremely critical, and this is the chief cause for the belief that this method is somewhat semi-quantitative. If the hydrofluoric acid residue pro-

cedure is to be used, evidently separate internal standards will have to be sought for strontium and for rubidium.

Potassium and ribidium usually behave very sympathetically in the arc, and Figure 10 shows a plot of I Rb 4202/I K 4047 vs. number of analyses, for the

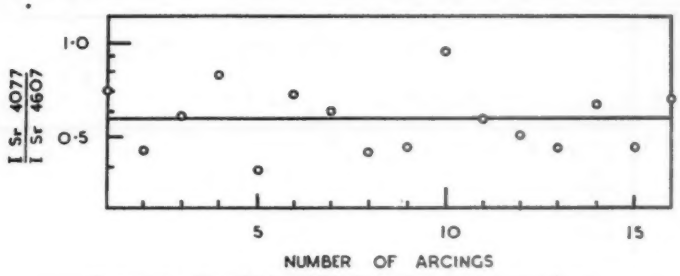


FIGURE 9.—*Plot of I Sr 4077/I Sr 4607 vs. number of arcings*

Spread of points about the mean very marked.

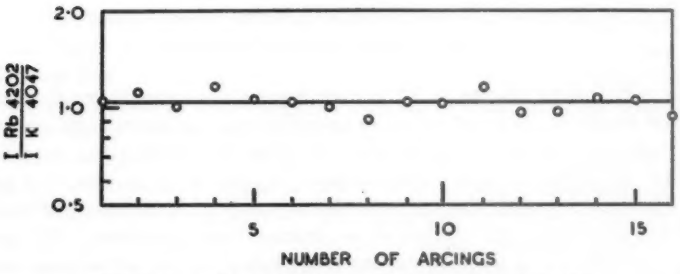


FIGURE 10.—*Plot of I Rb 4202/I K 4047 vs. number of arcings*

Reproducibility is good.

specimen of lepidolite already referred to. The standard deviation is about ±6%. A similar precision in the reproducibility of I Ca 4226/I Sr 4607 and I Ca 3933/I Sr 4077 has been observed by the author.

Although the potassium content of lepidolites is fairly constant, the variation is too great for the purposes of internal standardization, which is indeed unfortunate. If analyses of potassium in lepidolite by a single analyst are examined, for example, those of Stevens (1938), so as not to introduce personal error, the extreme variation is found to be about 15%, that is, about ±7.5% from the mean. Calcium, although present in lepidolites in only very small amounts, varies over relatively large ranges in concentration.

At this stage it is rather difficult to indicate the lines along which research on an accurate spectrochemical determination of Sr/Rb ratios should proceed. The spectrochemical procedure certainly does seem ideally suited. Many other sources of excitation could be tried, and much investigation can still be done with the d.c. arc. The development of an accurate method for determining the Sr/Rb ratio remains one

of the most urgent requirements for the proper development of the strontium method for measuring geological time.

A further requirement in this age method, is the accurate determination of the proportion of radiogenic strontium (Sr^{87*}) in the total strontium in any mineral.

DETERMINATION OF RADIOGENIC STRONTIUM IN TOTAL STRONTIUM

Common strontium consists of four isotopes, Sr^{88} (82.6%), Sr^{87} (7.02%), Sr^{86} (9.86%) and Sr^{84} (0.54%); all radiogenic strontium is of mass 87 (Sr^{87*}). Methods for estimating the different isotopes of an element depend essentially on the fact that each isotope is of slightly different mass, and the common procedure for making an isotope analysis is by means of the mass spectrograph. Relative abundance measurements on the different isotopes of strontium can be made with precision and ease, and, according to Mattauch (1947), an accurate quantitative analysis may be made on as little as 0.3 mg. of strontium salt, using his method of photographic recording. This is indeed good news for the strontium method because the concentration of strontium rarely exceeds 0.01% in the rubidium—rich minerals; but for this mass spectrographic sensitivity, the strontium method would be severely restricted and cramped in its application. Furthermore, according to a private communication from Dr. A. Keith Brewer, qualitative tests (Brewer, 1938) on the isotope composition of the strontium in lepidolites which he analyzed were made without any chemical pre-enrichment.

Plate 1, which is reproduced here with the kind permission of Prof. Mattauch, shows typical mass spectra of strontium isolated from several rubidium-rich minerals. In each of the mass spectra, Sr^{87} is either predominant or a major isotope. In many minerals, naturally, the proportion of radiogenic strontium is extremely small, and proportions of radiogenic strontium range from nearly 100% radiogenic strontium in some minerals, to negligible proportions, that is, say, less than 0.001% in others. An important limiting factor of the strontium method will be the lowest proportion of radiogenic strontium to nonradiogenic strontium that it will be possible to determine accurately with the mass spectrograph. According to Mattauch (1947), if Sr^{87*} is predominant, the mass spectrographic analysis is made by determining the intensity ratio of lines 87/88, since Sr^{88} produces the strongest line in common strontium; if the amount of common strontium approaches that of radiogenic strontium, the ratio of the lines 87/86 may be determined as well. If common strontium is predominant, it should be easy to detect a small quantity of Sr^{87*} since in common strontium the line-intensity ratio 87/86 is near unity and the accuracy in determining line-intensity ratios by the photometric method as used by Mattauch is greatest if the value of the intensity ratio is near unity. Even if the proportion of radiogenic strontium is 1%, the value of the intensity ratio 87/86 would increase 15% over the value for common strontium. In the absence of further data, 1–100% will be regarded tentatively as a *minimum* useful range in the proportion of radiogenic strontium in a mineral that can be accurately determined mass spectrographically.

It is fortunate that Sr^{87} in common strontium is not more abundant, because had this isotope been a major one the determination of small proportions of radiogenic

strontium in the presence of an excess of nonradiogenic strontium would not have been so convenient. In this respect, it may be recalled that K^{40} is beta active, and decays, in part only, to Ca^{40}, and there has been some speculation as to whether this transmutation could be utilized for the determination of geological age, particularly in certain micas. But since Ca^{40} comprises about 97% of common calcium, correction is more difficult, and there seems at the moment to be little likelihood of developing a really satisfactory calcium method for determining geological age.

Two other methods of demonstrating the presence of different isotopes of an element are in the field of optical spectroscopy. First, in optical line spectra, many lines may be split into hyperfine components, that is, the line has a hyperfine structure which depends on two properties of the nucleus, its spin and its mass. Once the hyperfine components of a given line corresponding to particular isotopes have been identified, the relative abundance of the different isotopes may be measured by determining the relative intensities of some of the hyperfine components. Heyden and Kopfermann (1937) were able to show that strontium in a specimen of lepidolite analyzed by Mattauch was radiogenic. The method is sensitive, and Heyden and Kopfermann employed only a few mgs. of strontium salt, but quantitatively is probably not as suitable as the mass spectrograph.

Second, in molecular spectra it has been shown that a slight change in the mass of one of the components of the molecular emitters is capable of causing a measurable shift in certain band heads, due to a change in vibrational frequency. Provided a large dispersion is used and resolution sufficient, the presence of isotope band heads due to strontium isotopes Sr^{88}, Sr^{87} and Sr^{86} may be observed in certain emission molecular spectra produced by the emitter SrF (Ahrens, 1948b). Quantitative isotope abundance measurements on these three isotopes may be made with a fair degree of accuracy, though not as accurately as with the mass spectrograph, but the amount of strontium required for a quantitative isotope analysis of this type is rather excessive (at least 200 mgs.) because the band sequence that is used is relatively weak.

In addition to the above methods, the suggestion has been made by Prof. Robley D. Evans of the Physics Dept., M. I. T., in a private communication, that the nuclear excitation of Sr^{87} could perhaps be used for determining the abundance of this isotope. Sr^{87} alone among the strontium isotopes is converted into an isomeric metastable (excited) state when bombarded with suitable missiles. If the intensity of excitation is measured, the amount of Sr^{87} (total) could be determined after suitable calibration has been made with standards. The principle underlying this method is not one of a variation in mass, but of nuclear configuration. Some preliminary experiments have been carried out using a relatively weak source of X-radiation, but so far no response from Sr^{87} has been detected; it is hoped to continue further investigations using a more powerful source and different missiles.

PREPARATION OF PURE LEPIDOLITE

Since lepidolite is considered a hydrothermal replacement mineral in pegmatite, any of the replacement minerals could be associated with it, either as side by side

association or as actual inclusions. Small quantities of associated minerals will interfere only if they carry nonradiogenic strontium. This interference can usually be connected by means of the mass spectrograph, but foreign material should be reduced to a minimum.

Quartz, mica, and feldspar are commonly associated with lepidolite. The presence of quartz is of no consequence and muscovite, unless present in appreciable quantity, should introduce negligible amounts of nonradiogenic strontium; the presence of feldspar, albite in particular, is more serious. Minerals which carry a relatively high concentration of common strontium and would be the most dangerous contaminants are the calcium minerals calcite, apatite, and fluorite, each of which may be found in a pegmatite. These minerals may on occasion be found associated with lepidolite, but actual inclusions have not been reported in the literature, nor have I found any. W. T. Schaller (private communication) has pointed out that apatite and calcite have been reported in muscovite, and therefore these minerals might be included in lepidolite. He states too that in a thin section of lepidolite from Harding, New Mexico, cut from a specimen of abundant, large masses of apatite, no evidence of included apatite could be found.

In any case, each specimen of lepidolite should be examined microscopically prior to an age determination. The presence of the most dangerous contaminants, calcite, apatite, and fluorite, could be easily ascertained, as could the presence of almost any other mineral likely to be associated with lepidolite. Lepidolite is bi-axial negative, having $2V = 25–60°$, and has $\alpha = 1.56$, $\beta = 1.598$ and $\gamma = 1.605$. A uniaxial variety has, however, been reported from Londonderry, Western Australia (Hendricks, 1939, p. 764) and two different specimens from the same general area which I examined also proved to be uniaxial.

From lepidolite of coarse texture, it is usually possible to obtain pure material by hand sorting, if necessary with the aid of binoculars after some coarse grinding. If the mineral is finer and more intimately associated with gangue, more drastic crushing is necessary, in which case other means of concentrating the lepidolite are to be sought. The method used for preparing some of the pure specimens of lepidolite referred to in this paper has been described in detail by Norman and O'Mear (1941).

Each specimen was crushed in a percussion mortar to −40 mesh and a portion was then ground in an agate mortar in the presence of some of the flotation reagent, a mixture of D.P. 243, sulphuric acid, and potash alum. Extra grinding in the presence of the flotation reagent served to make the lepidolite slightly finer and to condition it for flotation. Flotation was carried out in a miniature cell of the pneumatic type. The −200 mesh fines were removed from the rougher concentrate as they were of lower grade, and the +200 material was refloated; the product of the second flotation was taken as the final product. Precautions against possible sources of contamination were taken at each step.

Whether obtained as a flotation concentrate or not, most specimens have been cleaned by standing in hot aqua-regia for several minutes and then rinsing with distilled water. The cleaned mineral was then treated with pure hydrofluoric acid in a platinum dish and twice taken to dryness; decomposition and conversion to fluorides

was then complete. The fluoride residue was carefully baked, removed from the platinum dish, and ground in an agate mortar to ensure complete homogeneity of the specimen; it was then ready for arcing.

A blank examination of the hydrofluoric acid failed to reveal the presence of strontium or rubidium, and a trace of lead was the only impurity that could be detected.

STRONTIUM AGE DETERMINATIONS

The analytical data of Strassmann and Walling (1938) which were used for an indirect determination of the disintegration constant (λ) of Rb^{87} may now be used for determining the age of the specimen of lepidolite from southeast Manitoba. This age is calculated as 1800×10^6 years.

Hahn *et al* (1943) mentions a strontium age of about 530×10^6 years for pollucite from the Varuträsk (Sweden) pegmatite. A later strontium age on microcline from this locality is given by Eklund (1946) and the value of 1700×10^6 years is considerably greater than the earlier determination. Eklund states that a lead age determination carried out by W. Wahl gave an age similar to the strontium age.

Ishibashi and Ishihara (1942) determined the age of lepidolite from Korea as 1020×10^6 years; unfortunately this paper has been available in abstract form only.

In all, 30 strontium age determinations on lepidolite have been made using the spectrochemical procedure outlined in this paper (Table 6). Some of the ages have been given before (Ahrens, 1947a), but the values are slightly different because the older value of the disintegration constant was used. No other strontium age determinations are known to me. Some age observations have also been made on lithium-rich muscovite, pollucite, and amazonite.

DISCUSSION OF STRONTIUM AGE DETERMINATIONS

PALA, CALIFORNIA

The lepidolites from this area are from pegmatites associated with the great Californian batholith (Nevadan) and are generally considered to be pre-late Cretaceous to late Jurassic. Larsen and Keevil (1942) gave a tentative age of about $120–135 \times 10^6$ years to the earlier phases of the batholithic intrusions. Hurley and Goodman (1943) give two helium ages on two specimens of magnetite, 120 and 130×10^6 years.

The strontium ages are in general agreement with these findings; one strontium age (150×10^6 years) is probably excessive, one is semi-quantitative, and the other two indicate an age of about 100×10^6 years. This value is, however, possibly somewhat less accurate than most other ages, because the working curve (Fig. 6) has had to be extrapolated furthest to obtain these ages. Undoubtedly, accurate strontium age measurements could be made on the lepidolites from this interesting area.

NEW ENGLAND AREA

Many of the granite bodies from New England with which the lepidolite-bearing pegmatites are associated, are considered to be Acadian, probably Devonian.

In this area perhaps, the strontium ages have been the most disappointing. Evidently, contamination by nonradiogenic strontium is appreciable in some instances.

TABLE 6.—*Strontium age determinations on lepidolite*

	Locality	$\frac{I\ Sr\ 4077}{I\ Rb\ 4202}$	Probable geological age	Age from Sr/Rb ratios
1	Pala, California	0.017 0.015 0.014 0.015 Av.	Pre-late Cretaceous to late Jurassic.	110×10^6 yrs. "
2	"	0.020 0.025 0.025 0.023 Av.	"	150 "
3	"	0.011 0.017 0.013 0.014 Av.	"	100 "
4	"	Semi-quantitative as background excessive.	"	70–140 "
5	Newry, Maine	0.041 0.035 0.049 0.051 0.044 Av.	Acadian	300 "
6	Norway, "	0.039 0.044 0.042 0.025 0.041 Av.	"	280 "
7	Topsham, "	0.026 0.029 0.025 0.027 Av.	"	200 "
8	Mt. Mica, "	0.031 0.032 0.028 0.030 Av.	"	240 "

TABLE 6—*Continued*

	Locality	$\frac{I\ Sr\ 4077}{I\ Rb\ 4202}$	Probable geological age	Age from Sr/Rb ratios
9	Buckfield, Maine	0.043 0.043 0.075* 0.043 Av.	Acadian	300×10^6 yrs.
10	Auburn, "	0.081 0.078 0.066 0.072 Av.	"	540 "
11	Haddam, Connecticut	0.031 0.043 0.100* 0.038 0.054 0.033 0.033 0.040 0.039 Av.	"	270 "
12	Middletown, "	0.072 0.070 0.046 0.063 Av.	"	450 "
13	Portland, "	0.070 0.074 0.074 0.072 Av.	"	540 "
14	Dixon, New Mexico	0.10 0.12 0.06* 0.11 Av.	Pre-Cambrian	800 "
15	Brown Derby, Colorado	0.13 0.13 0.06* 0.13 Av.	"	900 "
16	" "	0.13 0.08 0.11 0.11 Av.	"	800 "

TABLE 6—*Continued*

	Locality	$\frac{I\ Sr\ 4077}{I\ Rb\ 4202}$	Probable geological age	Age from Sr/Rb ratios
17	Black Hills, South Dakota	0.13 0.13 0.13 0.13 0.12 0.13 Av.	Pre-Cambrian	900×10^6
18	" "	0.14 0.13 0.09 0.08 0.09 0.12 0.12 0.11 Av.	"	800 "
19	" "	0.21 0.40* 0.21 0.21 Av.	"	1500 "
20	Okongava Ost 72, South West Africa	0.10 0.10 0.13 0.10 0.09 0.10 0.105 Av.	"	750 "
21	Karibib, South West Africa	0.12 0.10 0.07 0.10 Av.	"	700 "
22	Albrecht's Höhe Karibib, South West Africa	0.13 0.09 0.15 0.11 0.12 Av.	"	850 "
23	Warmbad, South West Africa	0.15 0.13 0.14 Av.	"	950 "

TABLE 6—*Continued*

	Locality	$\frac{I\ Sr\ 4077}{I\ Rb\ 4202}$	Probable geological age	Age from Sr/Rb ratios
24	Omaruru, South West Africa	0.16 0.19 0.16 0.18 0.17 Av.	Pre-Cambrian	1150×10^6
25	Kubuta, Swaziland	0.33 0.24 0.29 0.29 Av.	"	2000 "
26	Falcon Island, Lake of the Woods, South East Manitoba	0.28 0.38 0.26 0.31 Av.	"	2200 "
27	Silver Leaf Mine, along Winnipeg Riv., S. E. Manitoba	0.20 0.38 0.28 0.24 0.30 0.38 0.25 0.295 Av.	"	2100 "
28	" "	0.32 0.40 0.26 0.33 Av.	"	2350 "
29	Along Winnipeg River, S. E. Manitoba	0.32 0.37 0.27 0.32 Av.	"	2300 "
30	Ontario, Canada	0.07 0.10 0.15 0.11 Av.	"	800 "

*Intensity ratio determinations marked with an asterisk have been omitted from the computation of mean intensity ratios.

Several lead ages available on pegmatites from the Portland area, Connecticut agree in magnitude, 250–300 $\times$ 10^6 years. Holmes (1947a) summarizes the data, and gives 260 $\times$ 10^6 years as the most probable age. Of three strontium ages made, two are evidently in error, whereas the third age, 270 $\times$ 10^6 years (Haddam Neck Specimen), agrees with the lead age. Without an isotope analysis, the presence of primary strontium in significant quantities will be a chief source of error; therefore, where disagreement occurs, only the lowest age can be considered. The Haddam Neck specimen is of excellent quality, consisting of large flakes which could be easily selected by hand, and the calcium lines were extremely faint. This specimen has been analyzed eight times and consequently the experimental error is relatively small.

From Maine, six specimens have been anlyzed. The age of one (Auburn) is excessively high, but the ages of the other specimens appear to be of the correct magnitude, although the variation in age from 200–300 $\times$ 10^6 years is quite considerable. The age of 200 $\times$ 10^6 years for the specimen from Topsham seems low, but a lead age of 230 $\times$ 10^6 years was obtained (Gonyer, 1937) on uraninite from the same locality, and this lead age is the lowest recorded for the Maine pegmatites. The four other strontium ages are from 240–300 $\times$ 10^6 years, the mean being 280 $\times$ 10^6 years.

NEW MEXICO—COLORADO—BLACK HILLS OF SOUTH DAKOTA

A lead age from the New Mexico pegmatites (Muench, 1938) is of the magnitude 800–900 $\times$ 10^6 years; the strontium age of 800 $\times$ 10^6 years is in agreement with the lead age.

Lead and strontium ages also agree closely in the pre-Cambrian in Colorado. Lead ages of pegmatites associated with the Pike's Peak and Silver Plume granites are 1050 $\times$ 10^6 and 930 $\times$ 10^6 years respectively. The lepidolites are from pegmatites near Brown Derby which are considered (Eckel, 1933) probably to be associated with a granite closely resembling the Pike's Peak granite. The strontium ages for two specimens of lepidolite, 800 and 900 $\times$ 10^6 years, agree in magnitude with the lead ages given above.

Three lead ages are available from the Black Hills of South Dakota: one (Davis, 1926) of uraninite from the well known Bob Ingersoll claim (1350 $\times$ 10^6 years) and two by Taylor (1935) of zircons from the Harney Peak granite and the Game Lodge granite, the ages of which have been recalculated using Wickman's (1943) graph and are respectively 1350 and 1640 $\times$ 10^6 years. The pegmatite of the Bob Ingersoll claim is believed to be associated with the Harney Peak granite, and the agreement between the granite and associated pegmatite ages is apparently good.

Strontium ages have been made on three specimens from the Black Hills pegmatites. Two of the specimens are from the Bob Ingersoll claim. The exact locality of the third specimen is not known; it shows considerable shattering, macroscopically and microspically, and is much older than the other two specimens. Since this specimen is of poorer quality and since in the absence of isotope data the strontium age is sometimes too high, the ages of the Bob Ingersoll specimens are considered to be more reliable.

Although on a very coarse scale the mean strontium age of 850 × 10^6 years is of the same magnitude as the lead age of uraninite from the Bob Ingersoll claim, the difference is considerable. Consequently the age of one of the Bob Ingersoll specimens has been determined seven times but the results have been the same.

In some publications, the uraninite age of the Bob Ingersoll claim has been included in a list giving selected lead ages, probably because an atomic-weight determination was made on the lead. The inclusion is open to question, however, for mineralogical reasons. Lead ages are invariably made on uraninite, whose properties (presence of crystal faces, hardness, general appearance, etc.) should indicate the mineral to be relatively unaltered. F. Hess, examined the uraninite analyzed by Davis (1926) and states in Davis' paper (p. 203), "None of them (the uraninite crystals) showed outward form and all are somewhat altered to gummite and uranophane."

An atomic-weight determination on lead from material altered as this would be of little value unless the ratio of Pb^{207}/Pb^{206} could be established. Also high ages are invariably obtained from altered uraninites, probably because of the ease with which UO_3 forms uranyl and uranate compounds which may be relatively easily leached out. Nor does the apparent agreement between the uraninite and zircon ages hold much water, because the value of both ages is doubtful. According to Taylor (1935) 10-gram concentrates of heavy minerals were prepared from 300 grams of granite, and each concentrate contained about 1–2% zircon. Extremely small amounts of lead (0.0004%) and uranium (0.002%) had to be determined and some contamination by primary lead seems certain, since apart from the possible presence of sulphide lead, potassium minerals contain traces of lead and in granite the lead content frequently averages about 0.002–0.003%. It seems probable therefore, that both zircon and uraninite ages are too high, and that the Black Hills pegmatites may well be of about the same age as the pegmatites from New Mexico and Colorado.

SOUTHERN AFRICA

There is much pre-Cambrian Shield exposure in Southern Africa, but only one lead age has been made on the pre-Cambrian pegmatites associated with the granite—gneisses, locally referred to as the 'Old Granite.' According to Holmes (1934), the age of uraninite from Gordonia, Namaqualand, is about 850 × 10^6 years. (Holmes gives a value of 900 × 10^6 years, and the above slightly different value was obtained by using Wickman's (1943) graphs.) The specimens analyzed were stated to be reasonably fresh, and several Pb/U ratios were in agreement.

Five specimens of lepidolite from the 'Old Granite' of South West Africa were analyzed, and the ages range from 700 to 1150 × 10^6 years. Although the range in variation is fairly large, one may conclude that the uraninite age is in general agreement with the strontium age; however, although the lepidolites and uraninite are all from the 'Old Granite,' they are separated by a very large distance.

The age of a specimen of lepidolite from Kubuta, Swaziland (southeast Africa) is of a different magnitude and more than twice the mean age of the South West African specimens. The Kubuta specimen is from an area which has been described by Hamilton (1938) and according to Dr. Hamilton (private communication) two 'Old

Granite' types can be recognized in this area, as well as a younger post-Witwatersrand granite. The lepidolite is from a pegmatite which is considered to be associated with the more ancient gneissic granodioritic 'Old Granite'. The presence of two 'Old Granites' has also been reported from Natal and Zululand.

Although the age of the Kubuta specimen is so very much greater than that of the other 'Old Granite' specimens, it is not necessarily high. Since only one specimen has been anzlyzed, however, no conclusions can be drawn and further comment is deferred until it is possible to collect more material and make isotope analyses.

ONTARIO

The strontium age of only one specimen was made, and it is in general agreement with the lead age measurements made on pre-Cambrian pegmatites from the Ontario-Quebec area.

SOUTHEAST MANITOBA

An inspection of available lead ages shows that in two areas on the earth's crust the pegmatite ages may be of a greater magnitude than for any other pre-Cambrian pegmatites; these areas are located in Northern Karelia and Southeast Manitoba.

The geological age of pegmatites from northern Karelia is quite often held to be the greatest age so far determined. Several specimens of uraninite have been analyzed (Khlopin and Vladimirova, 1938) and the ages which have been recalculated here with the aid of Wickman's graphs almost invariably equal 1800×10^6 years. Some specimens of monazite have also been analyzed (Khlopin and Vladimirova, 1938) and although the age of each specimen is about the same (mean 2100×10^6 years) it is somewhat higher than the uraninite age. From these data, the probable age of the pegmatites from northern Karelia is $1800–2100 \times 10^6$ years. Unfortunately, since no isotope analyses of the lead in these specimens are known, it has not been possible to compute ages according to Pb^{206}/U^{208}, Pb^{208}/Th^{232} and Pb^{207}/Pb^{206} ratios, which would provide valuable evidence for, or against, the above ages for the N. Karelian pegmatites.

Two specimens of uraninite and two of monazite from southeast Manitoba pegmatites have also given high ages, which, recalculated, range from $1700–2000 \times 10^6$ years. Lead from one specimen of uraninite and one of monazite has been analyzed on the mass spectrograph and serious discrepancies have been found in the ages, depending on the method of calculation (Table 7).

Nier (1941, p. 114) has pointed out that the Pb^{207}/Pb^{206} ages are quite likely the most reliable, and the agreement between the Pb^{207}/Pb^{206} uraninite and monazite ages is reasonably good. Consequently, although serious discrepancies are apparent, the agreement between these two ages indicates quite strongly that the age of the pegmatites from southeast Manitoba may be one of the greatest yet reported. Age determinations by the strontium method support this evidence.

First, if the analysis of the Silver Leaf Mine lepidolite by Strassmann and Walling (1938) is used, which analysis was utilized for the indirect determination of the disintegration constant of Rb^{87}, the age of the lepidolite can be calculated. 0.155% Sr

(> 99.7% of which is radiogenic) and 2.6% rubidium was found. The value for rubidium was considered to be somewhat approximate, but the authors believe the magnitude to be correct to within reasonably narrow limits. Using these data, an age of 1800 × 10^6 years is obtained for the lepidolite. According to Strassmann and Walling (1938), however, some strontium, roughly 10%, may have been lost during

TABLE 7.—*Lead ages of two specimens by different methods of calculation*
(Nier, 1938; 1941)

Mineral	Pb^{206}/U^{238} age	Pb^{208}/Th^{232} age	Pb^{207}/Pb^{206} age
Uraninite	1570 × 10^6 years	1252 × 10^6 years	2200 × 10^6 years
Monazite	3180 × 10^6 years	1830 × 10^6 years	2570 × 10^6 years

chemical manipulations; allowing for this loss makes the strontium age nearly 2000 × 10^6 years.

The ages of four separate specimens of lepidolite reported in this paper agree satisfactorily and tally reasonably well with the other ages in the southeast Manitoba area so far discussed. The strontium ages of these four specimens are from 2100 to 2350 × 10^6 years. Therefore, evidence seems now sufficient to establish beyond reasonable doubt what had been indicated by the lead method: that the pegmatites and associated granites from the southeast Manitoba area are of extreme antiquity and that, with the possible exception of the north Karelian pegmatites which may be of the same magnitude in age, and possibly also the Kubuta pegmatite, they represent the most ancient portion of the earth's crust on which it has been possible to determine directly the absolute age. Taking into consideration all the available age data on these pegmatites, an age of 2100 × 10^6 years is considered a reasonable value.

The granites from southeast Manitoba with which the pegmatites are associated, are intrusive into an ancient complex, referred to by Wright (1932, p. 12) as the Rice Lake Series, which is composed of various types of lavas, sediments—some of which contain pebbles of a still more ancient granite—and metamorphosed equivalents. The history of the series appears to have been quite extended. Consequently, some support is given to Holmes' (1947b) estimate of 3350 × 10^6 years as the age of the earth, rather than a younger age, that is, of about 2000 × 10^6 years which is sometimes considered to be the most probable age.

AGE OBSERVATIONS ON OTHER MINERALS

LITHIUM MUSCOVITE

The 'ages' of two specimens of lithium-rich muscovite from the 'Old Granite' of Southern Africa have been made. These ages of the same magnitude as the lepidolite ages, are: 1050 × 10^6 years for a specimen from Usakos, near the Omaruru-Karibib area in South West Africa, and 1200 × 10^6 years for a specimen from the Game Reserve, Eastern Transvaal. In the absence of sufficient research on the use of these minerals for age determinations, the above values do not have much practical value for actual age measurements; they do show however, that the proportion of

radiogenic strontium is predominant and that lithium-rich muscovite could very probably be used for making strontium age determinations.

Another specimen of lithium-rich muscovite was analyzed from the Black Hills, South Dakota. This specimen is sulphur yellow and had been labelled lepidolite but on analysis revealed a low lithia content, namely 0.2–1.0% Li_2O. The 'age' of the specimen was determined as 1600×10^6 years, and if the age of the pegmatite is assumed to be about 850×10^6 years, about 50% of the strontium can be regarded as radiogenic.

AMAZONITE

One specimen of amazonite, from Pike's Peak, Colorado, which contained 1.0% Rb, was analyzed and the 'age' was calculated as 1100×10^6 years. This age is similar to the lead age and supports the evidence of the mass spectrographic analyses of the Skuleboda amazonite and the indirect data on the Ilmen amazonites that strontium in amazonite is likely to be largely radiogenic.

POLLUCITE

Observations have been made on four specimens of pollucite: Karibib, South West Africa; Tin Mountain, Black Hills, South Dakota; Greenwood, Maine; and Varuträsk, Sweden. Only very approximate results could be obtained because background emission was found to be very intense, but in each case the approximate ages appeared to be about 2–3 times less than the expected age. Although no detailed investigations of the cause were made, the lowering of the pollucite age is quite probably due mainly to compositional changes: the presence of much cesium vapor in the arc column is bound to influence the intensity ratio I Sr 4077/I Rb 4202, which is extremely sensitive to slight changes of excitation conditions in the arc. This factor has been discussed and some data have been given in Ahrens (1947a).

EXTENDING THE SCOPE OF THE STRONTIUM METHOD

The usefulness of the strontium method would be increased and its scope extended if age determinations could be made on a greater variety of minerals and on minerals of wider distribution. There is little doubt that lepidolite is suitable for age determinations, and it is highly likely that age determinations could also be made on hydrothermal microcline, amazonite, and pollucite, but all these minerals are somewhat rare. Pegmatitic muscovite and biotite, in particular the lithium-rich varieties, are certainly worth investigating and since they are relatively common in pegmatite their use would be of considerable value. They are, however, restricted to pegmatites.

The application of the strontium method to rocks would increase its usefulness tremendously. At first thought, this possibility seemed rather remote because of a high concentration of common strontium relative to rubidium and radiogenic strontium in rocks. A closer scrutiny of the available analytical data provides cause for optimism, however; certainly basic rocks may be ruled out, but in some granite types which invariably contain 0.02–0.1% Rb the proportion of common strontium may be sufficiently low to permit a satisfactory quantitative analysis for the determination

of radiogenic strontium. One to 100% radiogenic strontium has been considered as a minimum, but tentative, useful range; for the purposes of this discussion, therefore, only those minerals or rocks which contain 1% or more radiogenic strontium will be considered.

Using analytical data published by Sahama (1945), the proportion of radiogenic strontium in some Finnish pre-Cambrian rocks was found (Ahrens, 1948a) to be quite considerable in some instances. Most of the granite and gneissic types were calculated to contain more than 1% and the proportion of radiogenic strontium in some was as high as 30%. Detailed data may be obtained from Ahrens (1948a), but the general conclusion may be made that the proportion of radiogenic strontium in these rocks should be sufficient to permit strontium age measurements.

The Caledonian rocks of Great Britain analyzed by Nockolds and Mitchell (1948) were found to contain an extremely low proportion of radiogenic strontium (0.01%) as a whole. In the biotite separated from the granite, the proportion of radiogenic strontium usually fell within 1–5%. Consequently it should be possible to determine the ages of these specimens by an analysis of the biotite fractions. These rocks are not very ancient (approximately 350×10^6 years) and hence the possibility of having a relatively high proportion of radiogenic strontium is less than in older rocks.

It would seem reasonable then that the strontium method could be applied to some granite-like rocks, in particular if the mica fraction could be separated, because the proportion of radiogenic strontium is likely to be at a maximum in the mica.

RADIOGENESIS OF STRONTIUM AND THE STABILITY OF THE SILICATE LATTICE

In silicate minerals where Rb^+ substitutes for K^+, it occupies a lattice site of high co-ordination. When Rb^{87} decays to Sr^{++}, an ion forms which (1) is slightly smaller than the parent ion, and (2) carries two charges instead of the single charge of the parent. The change in dimensions of the ions should produce negligible effects in the lattice because Sr^{++} is very similar in size to K^+ and only a relatively small number of Sr^{++} ions will be produced. The presence of a doubly charged ion in the place of the singly charged ion could cause some local electrostatic headaches, but it is very doubtful that any significant strain could develop within the lattice as a result of the formation of a limited number of doubly charged cations. Even in the most ancient minerals (2000×10^6 years), only about 0.6% of the total rubidium will have decayed, and in lepidolite, the mineral richest in rubidium and hence in radiogenic strontium, about 0.1–0.2% of the total number of K^+ sites will have been occupied by Sr^{++}: this proportion is extremely low and is undoubtedly insignificant. In the transmutation, $K^{40} \rightarrow Ca^{40}$, a similar situation obtains, and an extremely small proportion of K^+ sites are occupied by Ca^{++}.

Microscopic evidence has shown that in these ancient leipdolites, as well as in the younger, the mineral may be obtained in a fresh state of preservation and free from signs of internal strain. This condition may be contrasted with the internal strain and resultant poor state of uranium minerals, caused by the production of a large volume of helium, a large quantity of lead, and by the change of valence $U^{4+} \rightarrow U^{6+}$, on oxidation of UO_2 to UO_3: this change in valence results in a large change in size of the uranium ion. All these factors contribute to the production of shattering and to

the metamict state of uranium minerals, which makes them very susceptible to losing some of their constituents; whereas UO_2 is relatively inert, UO_3 more reactive, forms stable uranyl compounds in the presence of acidic solutions, and in the presence of alkaline solutions may form uranates. The shatter cracks in uranium minerals also form passage ways for possible losses of the gaseous products of decay—in particular, radon, whose half-life period is 3.85 days.

PRESENCE OF BARIUM AND OF CALCIUM AS A POSSIBLE AID FOR ESTABLISHING THE GENESIS OF STRONTIUM

If a mass spectrographic analysis of strontium in a mineral cannot be carried out, the concentration of either barium or calcium may sometimes serve as a useful aid in determining qualitatively, at least, whether the strontium is likely to be largely radiogenic.

It will be necessary to compare very briefly the geochemical characteristics of these three alkaline earths. Chemically they are similar, but the radii of their ions are somewhat different: $Ba^{++} = 1.43$ A, $Sr^{++} = 1.27$ A, and $Ca^{++} = 1.06$ A. Sr^{++} is of intermediate size and in a sense one may regard the geochemistry of strontium as intermediate between that of barium and calcium.

Strontium is invariably associated with calcium in calcium-rich minerals, but it may be present to some limited extent in potassium minerals, with many of which small amounts of barium are frequently present. Consider first the Ba:Sr association. In general, strontium and barium tend to concentrate in the potassium minerals of earlier formation, and residual hydrothermal solutions should contain only very small quantities of these two elements; furthermore, because the radius of $Ba^{++} > Sr^{++}$, the general tendency will likely be for barium to concentrate relative to strontium in the residual differentiates. For the Ca:Sr association, the reverse is probably true, and strontium tends to concentrate relative to calcium in the crystals of later formation. In this manner, strontium can be considered to hold an intermediate 'geochemical' position between calcium and barium.

More than a trace of calcium is rarely found in late potassium hydrothermal minerals, but should the supply of calcium not be exhausted, some may enter these minerals and will undoubtedly carry some primary strontium with it. The calcium contents of lepidolites are invariably very low, as evidenced by published chemical analyses, by Stevens (1938) for example, and also by personal observations on calcium spectra of the specimens so far analyzed. In a few instances, however, the calcium lines have been found to be unusually strong and far in excess of an amount which could have been generated by the radioactivity of potassium. In most, though not all, of these instances, the strontium ages have tended to be high. In no instance where the calcium spectra have been extremely faint, has the strontium age appeared to be excessive.

Some observations have been made on the barium spectra on ten of the specimens of lepidolite referred to in Table 6. In one specimen (No. 23) about 0.003% Ba was found, and in another (No. 24) about 0.0001–2 % Ba was found; in the eight other specimens (Ahrens, 1947a, no barium could be detected and is consequently less than about 0.0001 % in each case. Specimens No. 23 and 24 have a greater apparent age

than the other three specimens of lepidolite that have been analyzed from South West Africa. The spectra of the ten specimens were recorded on the quartz spectrograph and at the particular setting used observations could be made on the sensitive barium lines; the other spectra were recorded on the grating instrument, and the setting used did not include the wave-lengths of any very sensitive barium lines.

Naturally, this indirect evidence is of a very qualitative nature, but nevertheless I have found it of some value for the age determinations on the lepidolite specimens. It seems indeed likely that if the spectra of calcium are very faint and that if those of barium are absent, then the chances of a significant amount of contamination should be reduced to a minimum. Since calcium and barium, as well as strontium, are spectrally very sensitive, extremely small amounts of these elements can be detected. An experienced spectrochemist can readily estimate approximately the magnitudes of calcium and barium present by an inspection of the strengths of the lines.

BRIEF COMPARISON OF THE LEAD, HELIUM, AND STRONTIUM METHODS

EXPLANATION

As three methods which utilize the natural radioactivities of elements are now in a reasonable degree of development for the measurement of geological time, it seems appropriate to attempt a general comparison. Although factual data are still very scant, the brief discussion to be given is considered to be reasonably accurate and without bias and should portray, in a general way and rather tentatively, the relative merits of the lead, helium, and strontium methods.

APPLICATION

Lead method: Has been applied essentially to age measurements on pegmatites; satisfactory use of zircons would increase range of application very considerably.

Helium method: Suitable magnetite is relatively widely distributed and it appears that this method has potenialities of wide application.

Strontium method: So far confined to age measurements on pegmatites, but application to some igneous rocks possible, for example by determining the ages of biotite in granite.

USEFUL AGE RANGE

Lead method: Recent as well as pre-Cambrian. For very ancient ages, probably less reliable.

Helium method: Has been applied to fairly recent as well as pre-Cambrian measurements. In general, the more ancient pre-Cambrian measurements may be open to considerable hazard.

Strontium method: Not applicable to recent determinations, $<$ about 50–100 $\times$ 10^6 years. Best suited for pre-Cambrian and is probably the most reliable of the three methods for measuring very early pre-Cambrian time.

MINERALS GENERALLY USED AND STRUCTURE STABILITIES

Lead method: Uraninite, monazite, samarskite, and other strongly radioactive minerals. Lattice structure invariably disrupted and metamict condition common; serious consequences may result.

Helium method: Most satisfactory measurements thus far have been made on magnetite; satisfactory specimens can, according to present data, quite frequently be obtained.

Strontium method: Most work so far on lepidolite, which mineral can be obtained in an excellent condition. Other minerals such as hydrothermal pegmatitic microcline, amazonite, pollucite, Li-rich muscovite, which can very probably be used, can frequently be obtained in a satisfactory state.

POSSIBILITY OF A LOSS OF EITHER PARENT OR DAUGHTER ELEMENTS

Lead method: Chiefly as a result of poor state of highly radioactive minerals, chances of losses considerable; in particular, loss of U from minerals containing U as UO_3. Loss of Rn also possible.

Helium method: Since He is a gas, leakage is potentially great, but apparently He losses are insignificant in a high proportion of magnetites which show no evidence of recrystallization.

Strontium method: In lepidolite, and in the other minerals referred to above, parent and daughter elements should be tightly held by ionic bonding, and losses are considered quite negligible in good material. Unlike the helium method in particular, even a slight degree of recrystallization may have no serious consequences.

CORRECTION FOR THE PRESENCE OF SOME PRIMARY DAUGHTER ELEMENT

Lead method: Correction can usually be made for the presence of primary lead by means of the mass spectrograph.

Helium method: No direct correction can be applied to the entrapment of some He during the formation of a mineral, but indirect investigations thus far made have shown that, in the types investigated, little significant entrapment of He could be observed.

Strontium method: The presence of primary strontium can be corrected very easily on the mass spectrograph, even when the common strontium is in considerable excess; only about 0.3 mgs. of Sr salt are required.

METHOD, ACCURACY, AND SPEED OF ANALYSIS

Lead method: Pu, Th, and Pb usually analyzed by chemical methods, and unless these elements are present in very small quantities, accuracy is of a high order. Complete analysis of specimen rather tedious.

Helium method: Special He apparatus and suitable counting equipment. In most instances reproducibility about 10–15% per single determination. About 2–3 determinations may be made in a day.

Strontium method: Spectrochemical analysis probably best suited, speed considerable—10–15 age determinations can be made in quadruplicate in about 3 days—but the reproducibility thus far has been relatively poor; quadruplicate determinations can usually be reproduced to within 10–15%.

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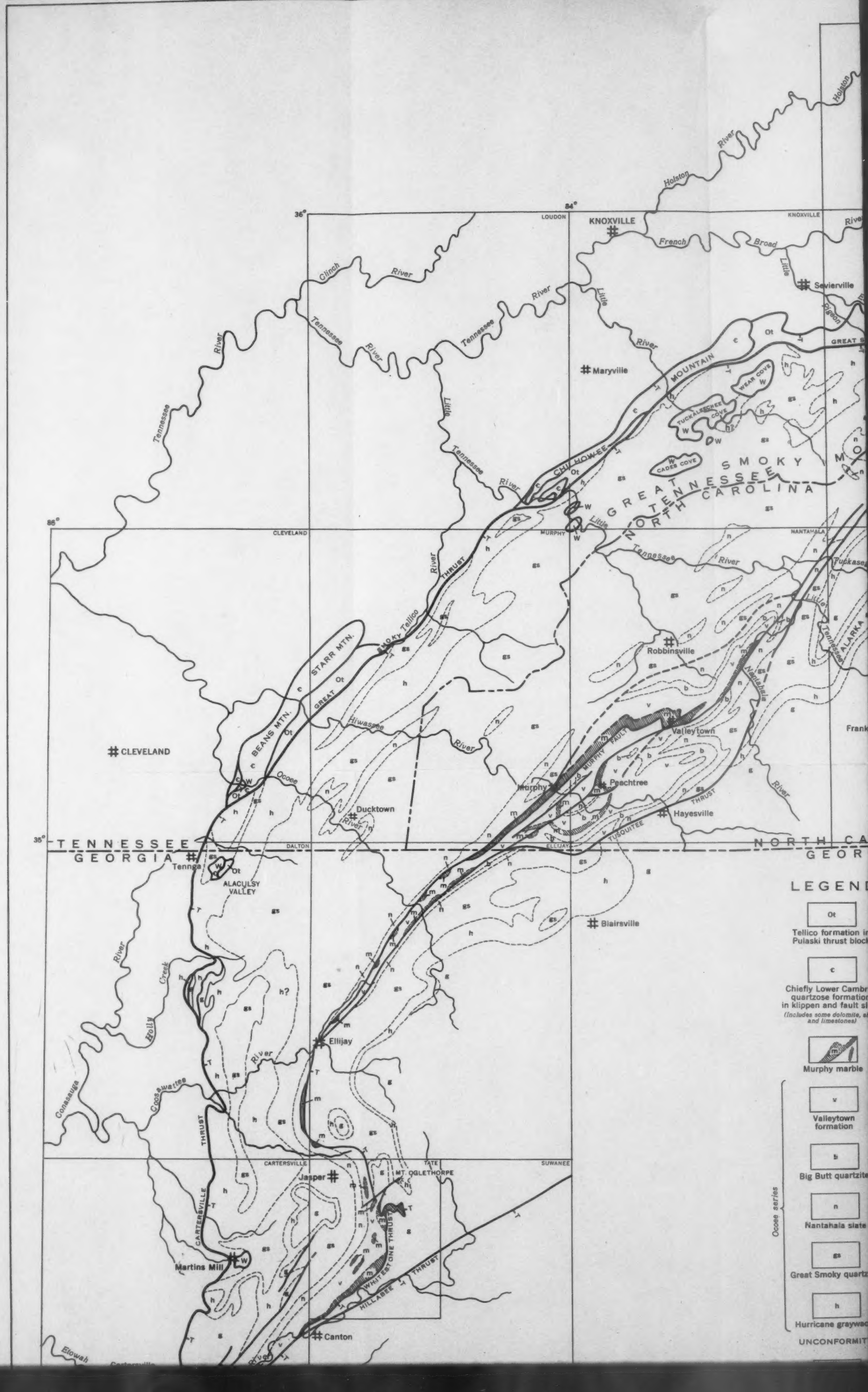
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WEAR COVE
TUCKALEECHEE COVE
CADES COVE
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NORTH CAROLINA
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MURPHY
NANTAHALA
Little Tennessee River
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Ocoee River
STARR MTN.
BEANS MTN.
GREAT
Hiwassee River
Nantahala River
Robbinsville
Valleytown
MURPHY FAULT
Peachtree
Murphy
Hayesville
TUSQUITEE
Ducktown
CLEVELAND
TENNESSEE
GEORGIA
NORTH CA
GEOR
DALTON
ELLIJAY
Tenn
ALACULSY VALLEY
Blairsville
Holly Creek
Conasauga River
Coosawattee River
Ellijay
CARTERSVILLE
SUWANEE
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Jasper
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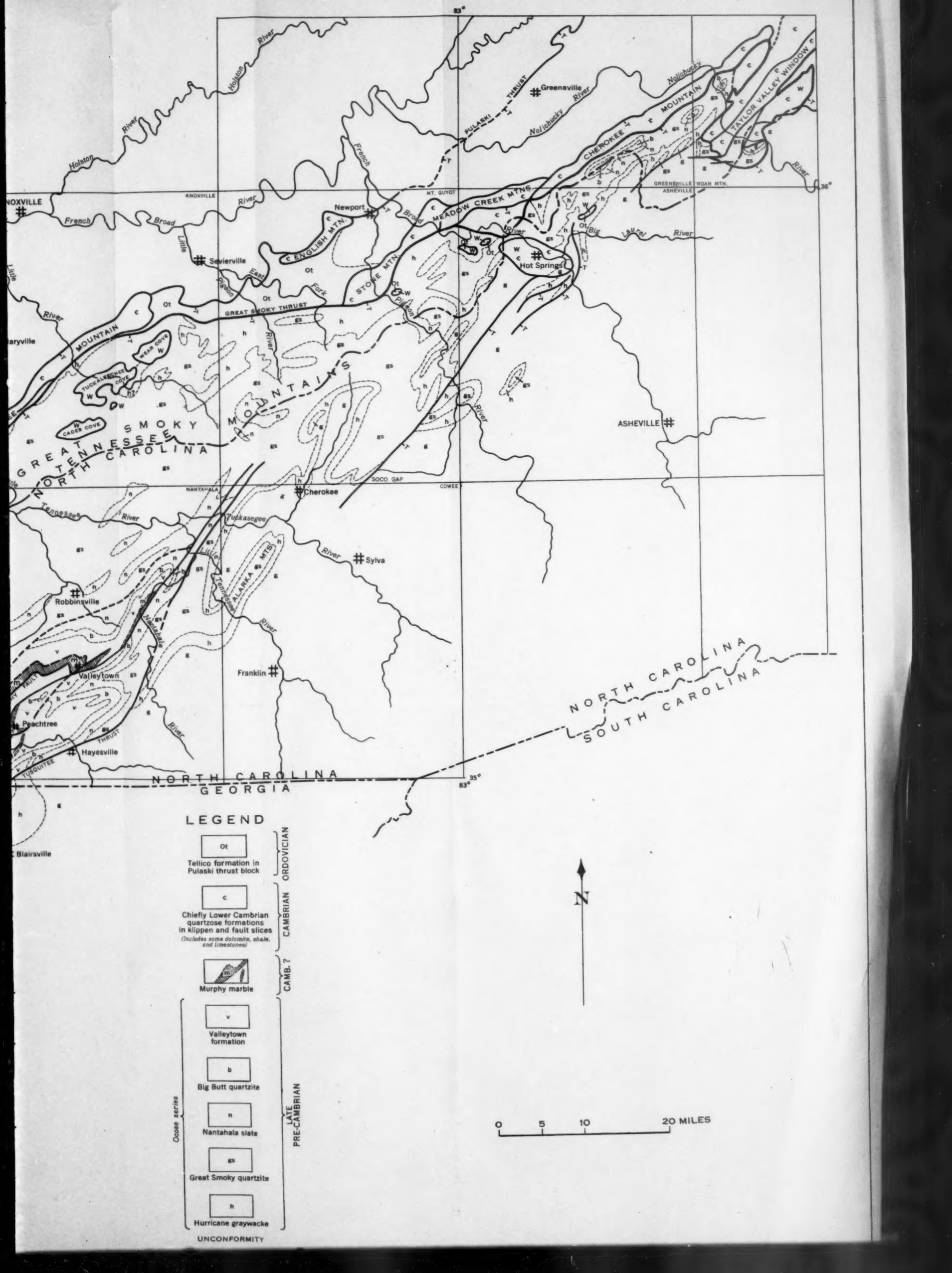
LEGEND
Ot
Tellico formation in Pulaski thrust block
c
Chiefly Lower Cambri quartzose formation in klippen and fault sl
(Includes some dolomite, sl and limestones)
m
Murphy marble
v
Valleytown formation
b
Big Butt quartzite
n
Nantahala slate
gs
Great Smoky quartzi
h
Hurricane graywac
Ocoee series
UNCONFORMITY



LEGEND

Ot	Tellico formation in Pulaski thrust block	ORDOVICIAN
c	Chiefly Lower Cambrian quartzose formations in klippen and fault slices *(Includes some dolomite, shale, and limestones)*	CAMBRIAN
m	Murphy marble	CAMB.?
v	Valleytown formation	LATE PRE-CAMBRIAN
b	Big Butt quartzite	Ocoee series
n	Nantahala slate	
gs	Great Smoky quartzite	
h	Hurricane graywacke	

UNCONFORMITY

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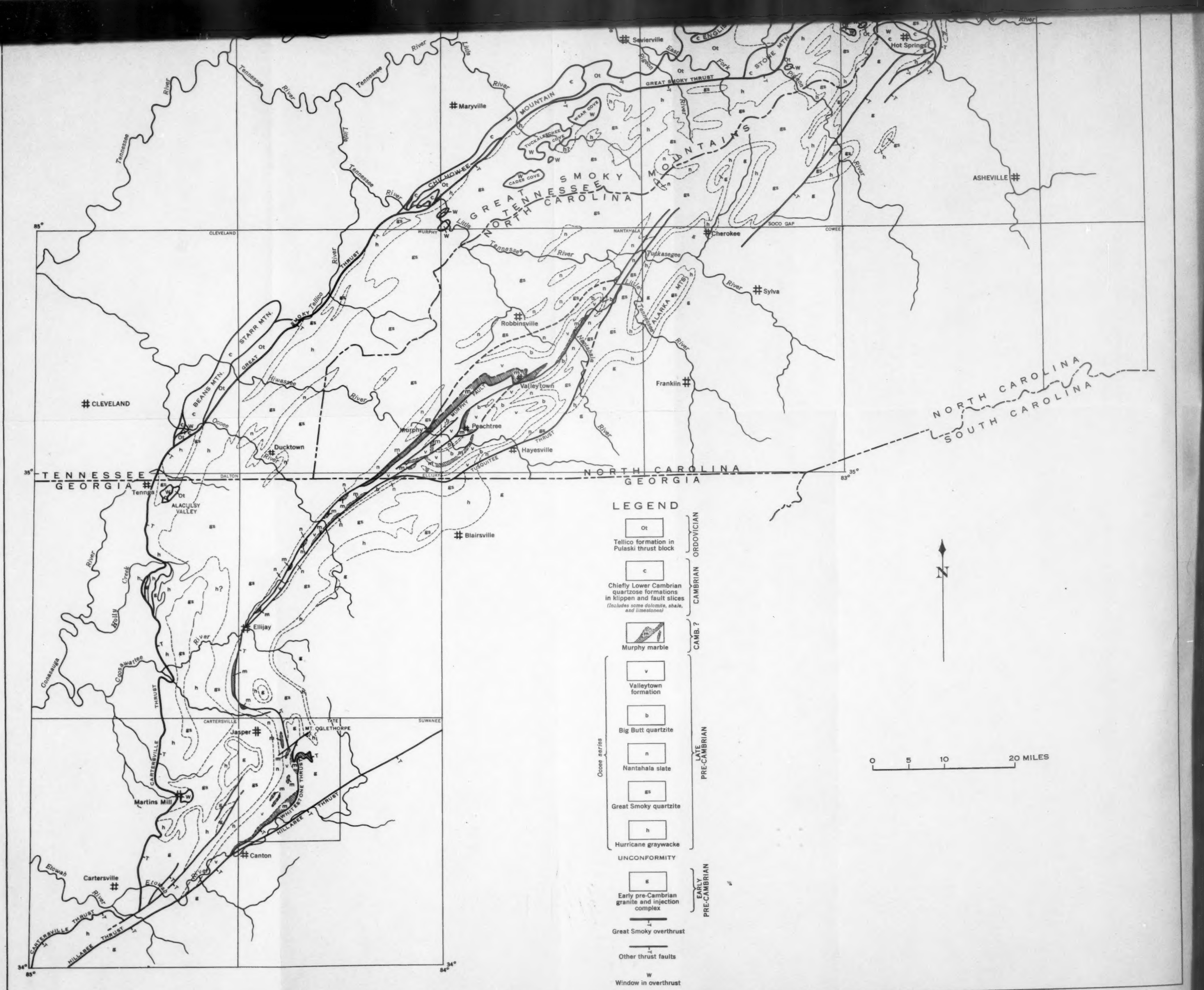

LEGEND

- Ot — Tellico formation in Pulaski thrust block — ORDOVICIAN
- c — Chiefly Lower Cambrian quartzose formations in klippen and fault slices (Includes some dolomite, shale, and limestones) — CAMBRIAN
- m — Murphy marble — CAMB. ?
- Ocoee series:
  - v — Valleytown formation
  - b — Big Butt quartzite
  - n — Nantahala slate
  - gs — Great Smoky quartzite
  - h — Hurricane graywacke
- LATE PRE-CAMBRIAN
- UNCONFORMITY
- g — Early pre-Cambrian granite and injection complex — EARLY PRE-CAMBRIAN
- Great Smoky overthrust
- Other thrust faults
- W — Window in overthrust

N

0 5 10 20 MILES

GENERALIZED GEOLOGICAL MAP SHOWING DISTRIBUTION OF THE OCOEE SERIES

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OCOEE SERIES OF THE SOUTHERN APPALACHIANS

BY GEORGE W. AND ANNA J. STOSE

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ABSTRACT

The Ocoee series is divided into four major units present from northern North Carolina to western Georgia and a fifth, younger, formation present only in southern North Carolina and Georgia. The units recognized are as follows:

Ocoee series, Late pre-Cambrian age	Valleytown formation
	Big Butt quartzite
	Nantahala slate
	Great Smoky quartzite
	Hurricane graywacke
	~~Unconformity~~~
Early pre-Cambrian	Injection complex

The Ocoee series is a broad synclinal belt of resistant Great Smoky quartzite, which forms most of the Great Smoky Mountains and the Bald Mountains, with Nantahala slate and in places the Big Butt quartzite enclosed in synclines; the Hurricane graywacke is on the flanks of the main syncline. In the deeper Murphy syncline the Big Butt quartzite is overlain by the Valleytown formation. The Ocoee series is thrust northwestward on the Great Smoky overthrust over Paleozoic rocks of the Great Valley, and in Tennessee overrides unmetamorphosed sandstone, shale, limestone, and limestone conglomerate which previously were mapped as Wilhite slate and were included in the Ocoee series. These rocks in a few places contain Middle Ordovician fossils and are a newly recognized part of a clastic shore facies equivalent to an expanded Tellico sandstone of undetermined stratigraphic range. This part of the Ordovician shore facies is in the Pulaski block in Tennessee and is exposed also in windows in the Great Smoky overthrust block in Tennessee, North Carolina, and Georgia. The Ocoee series does not resemble the Lower Cambrian Chilhowee group, which it overrides, nor any other Lower Cambrian facies. Lower Cambrian quartzites overlie with erosional unconformity the Catoctin basalt, Swift Run tuff, and Mt. Rogers volcanic series, proving that these volcanic rocks are late pre-Cambrian. The Ocoee series resembles the Lynchburg gneiss, which is equivalent to the late pre-Cambrian Swift Run tuff. The Ocoee series is stratigraphically overlain by the Lower Cambrian Unicoi formation and is late pre-Cambrian.

INTRODUCTORY STATEMENT

The writers (1944, p. 367–390, 401–416) defined and discussed the Ocoee series of the southern Appalachians and stated that the series is of late pre-Cambrian age, that it overlies the early pre-Cambrian injection complex, and is equivalent in age to several other rock series of different lithologic characters in other parts of the southern Appalachians. Recent field work confirming these conclusions was partly financed by a grant from the Penrose Bequest of The Geological Society of America.

The earlier paper (Stose and Stose, 1944, p. 369, 374–380) credited Safford (1856) with naming the Ocoee series from the type section along the narrows of the Ocoee River in southeastern Tennessee and assigning it to an Eozoic age. Keith in his early work in the Knoxville (1895b), Loudon (1896), and Mt. Guyot (unpublished) quadrangles, believed that the Ocoee series was of Silurian or later Paleozoic age but in the published reports on these areas he designated the series as of unknown age (Keith, 1895b, p. 2); while Hayes (1895, p. 2) in the Cleveland folio, which covers the type locality of the Ocoee, stated that the series was of probable Algonkian age. Keith (1904, 1907 c), in later work in the adjacent Asheville and Roan Mountain quadrangles, assigned rocks equivalent to the Ocoee series to the Lower Cambrian, abandoned the formation names used in earlier reports, and applied the same names to the formations of the Ocoee series that he gave to the Lower Cambrian quartzites of the Chilhowee group.

The writers agree with Hayes that the Ocoee series is of late pre-Cambrian (Algonkian) age for the following reasons: They have found that sediments of the Ocoee series do not resemble those of the Chilhowee group or other facies of Lower Cambrian quartzites either in lithologic character or manner of deposition; that the Ocoee series and the Chilhowee group originated in widely separated areas and that their present proximity is the result of overthrusting; that in Tennessee, in the gorge of Nolichucky River, basal beds of the Lower Cambrian Unicoi formation overlie the Ocoee series with a stratigraphic unconformity although without a marked angular discordance; that the Ocoee series overlies the early pre-Cambrian injection complex with a pronounced stratigraphic, structural, and metamorphic break and is much younger than the complex; that the Ocoee series lithologically resembles the Lynchburg gneiss of late pre-Cambrian age.

For the above reasons the writers place the Ocoee series in the late pre-Cambrian and will refer to all the rocks of the Ocoee series, including those to which Keith later gave Lower Cambrian formation names, as late pre-Cambrian. They place in the Ocoee series also some rocks previously mapped as part of the Carolina gneiss in the area east of the Whitestone fault in northern Georgia. They have excluded from the Ocoee series, as formerly mapped, unmetamorphosed limestones, shales, and sandstones, probably of Ordovician age, which are present in southeastern Tennessee and extend southward to northern Georgia. These rocks, here called the Tellico formation, are in a thrust block which the writers regard as the southwestern extension of the Pulaski over-thrust block of Virginia. They are overridden by the Great Smoky overthrust block and in places are exposed in windows in that block.

OCOEE SERIES (LATE PRE-CAMBRIAN)

LOCATION AND EXTENT

The Ocoee series forms the high mountains on the western border of the Blue Ridge Plateau in eastern Tennessee, western North Carolina, and northern Georgia (Pl. 1).[1] On their west side these mountain ranges face the Great Valley, a lowland largely underlain by Paleozoic limestones. At the northeast end of the Ocoee series the bordering range is called the Unaka Mountains; farther southwest, the Bald Mountains. Southwest of Pigeon River to Little Tennessee River they become the Great Smoky Mountains where many peaks more than 6000 feet in altitude culminate in Clingmans Dome, 6643 feet. South of Little Tennessee River the border range continues with diminishing elevation and terminates in northern Georgia in the Cohutta Mountains.

Southeast of the border ranges the Ocoee series forms lower mountains, such as the Alarka and Tusquitee, which grade eastward into the general surface of the Blue Ridge Plateau, a region largely underlain by the granitic gneisses of the injection complex. The plateau has an altitude from 3000 to 4000 feet and in places is deeply dissected. Many mountains rise above its surface and on the southeast side of the plateau reach an altitude of 6684 feet in Mt. Mitchell, the highest mountain in the eastern United States. The Blue Ridge escarpment bounds the plateau on the southeast and overlooks the Piedmont Upland, 1500 feet below. In North Carolina this escarpment lies 35 to 40 miles southeast of the Great Smoky Mountains. In Georgia the more westerly trend of the escarpment narrows the plateau, which ends at Mt. Oglethorpe.

The Ocoee series lies on the western border of a great overthrust block. Where this bordering fault crosses the Knoxville quadrangle Keith (1927) named it the Great Smoky thrust. The writers have mapped the extension of this overthrust in Tennessee (Pl. 1) and have traced it into the Cartersville overthrust in Georgia (Hayes, 1891) and Alabama. At the northeast end of the belt of the Ocoee series, it is a synclinal remnant of that fault block 4 miles wide (Stose and Stose, 1944, p. 384–385, Figs. 2A, 2B); the series extends southwestward with widening outcrop across northwestern North Carolina and southeastern Tennessee and passes into Georgia and Alabama. From western Georgia northeastward the Ocoee series maintains its synclinal character, the youngest beds being exposed only in the centers of synclines. At the northeast the major syncline follows the Bald Mountains (Fig. 5) and rises southwestward. In the Mt. Guyot quadrangle the syncline follows the trend of the main crest of the Great Smoky Mountains to the vicinity of Newfound Gap, then trends more southerly and passes southwestward into the Murphy syncline.

On the northwest side of the syncline the Hurricane graywacke, the oldest formation of the series, is present for the most part along the Great Smoky overthrust. On the southeast side of the syncline this basal formation lies unconformably on the

[1] Places to which references are made in this paper and are not shown on Plate 1 may be found on the topographic maps of quadrangles and map of the Great Smoky Mountain National Park of the U. S. Geological Survey and on maps of the U. S. Tennessee Valley Authority.

injection complex, a relationship observed by the writers at many places from northeastern Tennessee southward to Mt. Oglethorpe, Georgia. The basal contact of the Ocoee series is irregular in trend because the rocks are closely folded and in the northeastern part (Fig. 5) of the area the folds are cut by minor thrust faults within the Great Smoky overthrust block. In southern North Carolina and northern Georgia the southeastern part of the belt of the series in the vicinity of the Murphy syncline is broken by the Tusquitee, Murphy, and Whitestone overthrusts (La Forge and Phalen, 1913) which, in the neighborhood of Hayesville, and south of Tate, carry the injection complex over the Ocoee series. At Canton and southwestward across Georgia and Alabama the Ocoee series is overriden on its southeast side by a thrust fault which is called the Hillabee overthrust in Alabama.

SUBDIVISIONS

The writers have divided the Ocoee series into four major units which can be readily recognized and which persist throughout the extent of the series to Cartersville, Georgia. A fifth, and uppermost, formation occurs only in the Murphy syncline. No attempt was made to subdivide the Ocoee series southwest of Cartersville in Georgia and Alabama. The writers' mapping (Pl. 1) is based on traverses and not on detailed field work in the heavily wooded ranges of the region.

In the Asheville quadrangle (Fig. 5) the full sequence of the Ocoee series is not present because it is narrowed by thrust faulting and interrupted by the Hot Springs window (Stose and Stose, 1944, p. 385–386). Southwestward, where the belt of the series widens in the region of the Great Smoky Mountains, the section of the rocks in the southeastern part of the belt is not clear because there the rocks are highly metamorphosed and closely folded.

The type section proposed by the writers is taken from the Bald Mountains, Greeneville quadrangle (Fig. 5), because there the rocks are not greatly metamorphosed and, except for minor folding, the series rises progressively northward from its base to the youngest beds, which lie in a gentle syncline at the crest of the Bald Mountains.

The section with estimated thicknesses of the four formations of the Ocoee series observed on the Appalachian Trail from Devils Fork Gap to the Big Butt in the Bald Mountains of the Greeneville quadrangle is given in Table 1.

The new name Hurricane graywacke is here applied to the oldest formation of the Ocoee series because of its fine exposures on Little Hurricane Creek and at the Hurricane settlement, N. C. (Fig. 5), 6 miles northeast of Hot Springs, N. C. The initial sediments of graywacke in this vicinity and northeastward rest on granite gneiss of the injection complex. In the Asheville and several other folios Keith (1904; 1905a; 1907c; Mt. Guyot folio) used the name Snowbird formation for these beds which he assigned to the Lower Cambrian. He included part of the Great Smoky conglomerate in the Snowbird formation at the type locality (Snowbird Mountain in the Mt. Guyot quadrangle) and elsewhere. He (1904; 1907c) applied the name Snowbird also to well-established Lower Cambrian rocks. For these reasons the writers believe that the name Snowbird is misleading and have applied the new name (Hurri-

cane) to the graywacke which underlies the Great Smoky quartzite and is the basal formation of the Ocoee series.

The names Great Smoky conglomerate and Nantahala slate were used by Keith (1907a), La Forge and Phalen (1913), and Bayley (1928) in the Nantahala, Ellijay,

TABLE 1.—*Ocoee series (late pre-Cambrian), Bald Mountains, Tenn.-N. C.*

Formation name	Lithologic description	Feet
Big Butt quartzite. Exposed on Big Butt and Ball Ground (Murray shale and Nebo quartzite of Keith, 1905a)	Thick-bedded white granular quartzite and interbedded argillite and fine-grained greenish arkose	160±
Nantahala slate (Nichols slate of Keith, 1905a)	Black slate with some beds of fine-grained dark-green arkosic quartzite (poorly exposed)	200±
Great Smoky quartzite Exposed on Green Ridge, Snake Den Ridge, and Gravel Knob. (Cochran conglomerate and Hiwassee slate of Keith, 1905a)	White conglomeratic arkosic quartzite and interbedded black slate; coarse white quartz pebble conglomerate with pebbles and streaks of black fine-grained slaty graywacke and a thin metadiabase sill; porous-weathering arkose and thin conglomerate beds	500±–800±
Hurricane graywacke (Snowbird formation and Hiwassee slate of Keith, 1904; 1905a)	Slaty dark argillite; dark siliceous fine-grained graywacke; at base, black to dark-gray fine-grained graywacke containing angular grains of dark glassy quartz and feldspar	500–1000±
~~UNCONFORMITY~~~		
Injection complex (Cranberry granite and Max Patch granite of Keith, 1904; 1905a)	Granitic and dioritic gneisses, injection gneiss, and metasediments	

and Tate quadrangles, and the writers retain these names for the same units but they place them in the late pre-Cambrian and not in the Lower Cambrian as they were defined by Keith. The Great Smoky formation is essentially a dark grit containing angular grains of quartz, feldspar, and granite gneiss and might well be called a graywacke. Certain beds contain pebbles of quartz, feldspar, and granite gneiss, but the pebbles seldom exceed half an inch in diameter. The writers use the name Great Smoky quartzite, emphasizing its quartz-rich character. The writers have divided the Great Smoky quartzite into lower, middle, and upper members each of which has distinctive lithologic characteristics. The middle member contains most of the conglomeratic beds. These members are described in the sections but have been mapped only locally (Pl. 2).

The Big Butt quartzite is named (Stose and Stose, 1944, p. 404–405) from the Big Butt, the northeast culmination of the Bald Mountains (Fig. 5), where massive white

quartzite and interbedded green argillite and fine-grained arkosic quartzite are enclosed in the syncline. Keith (1905a) called the quartzite Nebo and the softer beds Murray shale, which are names that properly should be applied only to Lower Cam-

TABLE 2.—*Ocoee series in Great Smoky Mountains, Tenn.-N. C.*

	Proposed formation names	Names used in the Knoxville, Loudon, and Cleveland folios, and on the Mt. Guyot map. (Keith, 1895b–1896; Mt. Guyot folio; Hayes, 1895.)	
Late pre-Cambrian	Nantahala slate	Hazel slate (Pigeon slate included in places)	Age unknown
	Great Smoky quartzite	Clingman conglomerate, Thunderhead slate, Thunderhead conglomerate, Cades conglomerate, Pigeon slate (Citico conglomerate and Snowbird formation included in places)	
	Hurricane graywacke	Citico conglomerate and Wilhite slate, in part (Pigeon slate, Hiwassee slate, and part of the Snowbird formation included in places)	

TABLE 3.—*Ocoee series (late pre-Cambrian) in region southwest of Great Smoky Mountains*

	Proposed formation names	Names used in the Nantahala folio, Murphy map, Ellijay folio, and Tate report (Keith, 1907a, Murphy folio; LaForge and Phalen, 1913; Bayley, 1928)	
Late pre-Cambrian	Valleytown formation	Valleytown formation Brasstown formation	Lower Cambrian of Keith
	Big Butt quartzite	Tusquitee quartzite	
	Nantahala slate	Nantahala slate	
	Great Smoky quartzite	Great Smoky conglomerate	
	Hurricane graywacke	Not separated on the published maps. Where present, it is included in the Great Smoky conglomerate	

brian formations, hence the writers use a new name for the youngest formation of the Ocoee series in the Bald Mountains. The Valleytown formation (Keith, 1907a), the youngest formation of the series in the Murphy syncline, is not present in the Bald Mountains.

The major formational units recognized by the writers in the Great Smoky Mountains are given in Table 2.

The formations recognized by the writers in southeastern Tennessee, southwestern North Carolina, and northern Georgia, and those used in the published and unpublished folios of that region are given in Table 3.

HURRICANE GRAYWACKE

Definition.—The Hurricane graywacke, the basal formation of the Ocoee series, is a fine- to medium-grained grayish-black or grayish-green graywacke consisting principally of angular grains of glassy quartz and feldspar in a matrix of mica and black, iron-oxide dust. On the southeast side of the belt of the series the Hurricane rests unconformably on the early pre-Cambrian injection complex. On the northwestern side of the belt it is exposed in places along the Great Smoky overthrust. It occurs also in anticlines within the belt, and in Georgia it surrounds early pre-Cambrian rocks exposed in the Fort Mountain, Corbin, and Salem Church anticlines.

General distribution.—At the northeast end of the belt of the Ocoee series (Roan Mountain and Chestoa quadrangles) small areas of the Hurricane graywacke overlie the injection complex and are thrust over (Fig. 6) the Holston Mountain thrust block and the Taylor Valley window in that block (Stose and Stose, 1944, p. 382–383). The main southeastern belt of the graywacke extends from a point 3 miles southwest of Erwin, Tenn., passes through Hurricane, N. C. (Fig. 5), the type locality, and extends southwestward to northern Georgia and ends at Mt. Oglethorpe.

The Hurricane graywacke in this area overlies rocks of the injection complex, with which it is closely folded, and dips north or northwest under the Great Smoky quartzite. East of Hurricane, within the injection complex, the graywacke is enclosed in two narrow parallel synclines which are faulted and overridden on their east sides by the complex. These faults, and other parallel southwest-trending faults, which offset the border fault of the Hot Springs window (Stose and Stose, 1947, p. 636–637), are marked by strong dislocation metamorphism, shown by mylonitization of the overthrust granite gneiss. One of these thrusts marked by granite mylonite extends southwest and passes through Soco Gap (Pl. 1). The fault is within the injection complex except at a point 12 miles northeast of the gap where a syncline of Ocoee with basal beds of the Hurricane graywacke resting on the complex lies east of the fault and is thrust westward over the Great Smoky quartzite. North of Cherokee (Pl. 1) an area of granite extends northward along the valleys of Straight and Raven forks. The writers (1944, p. 411) stated previously that this granite gneiss intrudes the Ocoee series. Their recent field work indicates that the granite gneiss is part of the injection complex exposed in an anticline in the series and that the Hurricane graywacke stratigraphically overlies the granite gneiss. South of Bryson (Cowee quadrangle) the Hurricane surrounds a north-trending syncline of the Great Smoky quartzite.

The Hurricane graywacke north of Blairsville, Ga. (Pl. 1), is the basal part of the Ocoee series which was mapped previously as Carolina gneiss (La Forge and Phalen, 1913). The graywacke bends around the north-pitching end of the Mt. Oglethorpe syncline and is cut off by the Whitestone thrust at a point 3 miles north of Tate Station (Fig. 1).

On the northwest border of the belt of the Ocoee series, the Hurricane graywacke is in disconnected areas along the Great Smoky overthrust (Pl. 1). One of the larger areas extends from a point west of the Hot Springs window southwestward into the Knoxville quadrangle where, just west of West Fork of Little Pigeon River (Fig. 3),

it passes under the Great Smoky quartzite, which there lies at the front of the Great Smoky overthrust. To the southwest another large area of the Hurricane extends from the vicinity of Little Tennessee River to near Cartersville, Ga. This is in Safford's (1856) type section on Ocoee River, where the graywacke lies just east of the Great Smoky overthrust. In Georgia this belt of graywacke is narrowed and is cut out in places by overthrusting, namely just east of Tennga, south of the Holly Creek (Pl. 1) embayment of the Cartersville thrust, and east of Cartersville.

Lithology of southeastern belt.—The Hurricane graywacke from the northeast end of its area southwestward to the eastern border of the Mt. Guyout quadrangle is lithologically like the formation in the section in the Bald Mountains (Table 1). The sequence is introduced by hard, dark, dense, finely banded quartzite, containing grains of dark, glassy quartz and feldspar, with interbedded dense, black, graphitic, slaty graywacke containing thin gray current-bedded siliceous layers. Overlying are dark-gray to green, fine graywackes commonly without visible parting planes but in places with thin, dark-gray, siliceous partings and thin beds of grayish-green, impure quartzite. These quartzose beds grade upward into light-gray quartzites, basal beds of the Great Smoky quartzite. The Hurricane graywacke is dense, indurated, and cut by well-developed cleavage planes and, even where quartzose layers are present, the rock does not part along bedding planes but breaks easily into slabs along the cleavage.

The Hurricane graywacke from the east edge of the Mt. Guyot quadrangle southwestward to Georgia is composed of graphitic schist, commonly garnetiferous, muscovite-quartz schist, and thin dark quartzite with muscovite blades on parting planes. In northern Georgia the Hurricane graywacke includes graphitic schist, kyanite-bearing muscovite schist, and fine-grained arkosic quartzite, which at many places are intruded by mica pegmatite. Throughout this more-metamorphosed belt the writers (Pl. 1) have arbitrarily separated the Hurricane graywacke from the Great Smoky quartzite.

Lithology of the northwestern belt.—Massive, dark-green, fine-grained graywacke characterized by spheroidal weathering and overlying fine-grained quartzite make up most of the formation in this belt and the basal dark quartzite is exposed only at a few localities. The Hurricane is well exposed on Pigeon River, southeast of Hartford (Fig. 3), and on East and West forks of Little Pigeon River. On Ocoee River (Cleveland quadrangle) in the type section of the Ocoee series (Safford 1856), it crops out from the border of the Great Smoky overthrust at the mouth of Greasy Creek (Fig. 7) southwestward for 5 miles, and at the southeast dips 70° E. under the Great Smoky quartzite. The section on Ocoee River contains infolded black quartzite which resembles the basal beds of the Hurricane and may be such beds brought up in an anticline. From Safford's type locality fine-grained green graywacke of the Hurricane extends southwestward into Georgia. Southwest of Cartersville, Ga., and in Alabama the formation is a green phyllite. In Georgia, Hayes (Cartersville folio) called part of this phyllite the Gilmer schist, and Crickmay (1936) included it in the Talladega series. In Alabama it is mapped as the Talladega slate (Butts, 1926).

Lithology adjacent to thrust faults.—The Hurricane graywacke on the border of the Great Smoky overthrust is a crinkled, thin-banded, siliceous phyllite with a cleavage

that dips 30°–40° SE. parallel to the dip of the fault. The phyllite on that overthrust is exposed west of Mundy Gap (Fig. 5) where it rides over Lower Cambrian rocks of the Cherokee Mountain block; at a point about a mile sothwest of this gap where it is thrust on Shady dolomite (Stose and Stose, 1947, Fig. 2) of the Paint Rock slice; on U. S. Highway 25-70 northwest of the Hot Springs window (Fig. 5); just south of Cosby (Fig. 3); and at Greasy Creek (Fig. 7) in the type section of the Ocoee series.

In Metcalf Bottoms (Fig. 4) the graywacke in an anticline in the Great Smoky quartzite is exposed in the gorge of Little River. Three fourths of a mile northeast of The Sinks, the graywacke dips southwest beneath the Great Smoky quartzite in a syncline, the center of which passes through The Sinks. On the west side of the syncline sheared and slickensided phyllite, probably representing the lower beds of the Great Smoky quartzite and underlying Hurricane graywacke, form a belt 2 miles wide, which extends to the south edge of the Tuckaleechee Cove window. The phyllite in a rocky ledge called "Smoky Face" stands out boldly above Little River. These deformed rocks are near the sole of the Great Smoky overthrust, which is eroded through at Tuckaleechee Cove, exposing overridden Upper Cambrian or Lower Ordovician limestone. In places the phyllite is tightly crinkled; elsewhere it is thinly laminated, with low-dipping wavy cleavage intersected by a second cleavage so that it breaks readily into small slickensided rhombic blocks. At the contact of the overthrust graywacke with the limestone of Tuckaleechee Cove (Fig. 4) the fault dips 50° S. The graywacke at the fault is a shoepeg slate, so broken by the movement that no well-defined cleavage is discernible.

A similar phyllite of the Hurricane graywacke is present along the Cartersville overthrust in Georgia. At a point 5 miles north of Martins Mill (Pl. 1) the thrust plane dips 30° E., and the phyllite is in contact with, and overthrust on, blue argillaceous limestone of the Middle Cambrian Conasauga shale. In the vicinity of Martins Mill the phyllite, developed along the thrust, surrounds a small window of the Rome block exposing Weisner quartzite (a name locally applied to undivided Lower Cambrian quartzites) and overlying Shady dolomite and Rome formation. A minor thrust that extends southwestward from the northeast end of the Corbin granite and passes into the Cartersville thrust southeast of Cartersville cuts off the southeast side of the Corbin anticline and carries crumpled green phyllite, apparently of the Hurricane graywacke, over the Corbin granite. From this point the phyllite extends southwestward along the Cartersville overthrust across western Georgia and Alabama.

Hurricane graywacke surrounding granite gneiss exposed in anticlines.—In Georgia, south of Holly Creek (Pl. 1) and east and south of Martins Mill, three anticlines in the Ocoee series—the Fort Mountain, Salem Church, and Corbin—bring to the surface granite gneiss overlain by basal beds of the Ocoee series on the flanks of the folds. Previously the writers (1944, p. 411) accepted the conclusion of Bayley (1928, p. 48–49) that the Salem Church granite intrudes the Hiwassee slate of the Ocoee series and the statement of Crickmay (1936, p. 1384) that the Corbin granite is intrusive in the Talladega series which the writers include in the Ocoee series. The Salem Church granite, exposed in an oval area in the western part of the Tate quad-

rangle (Bayley 1928) and adjacent part of the Cartersville quadrangle (Pl. 1), is surrounded by a band of Hiwassee slate (basal formation of the Ocoee series) and several small areas of the slate lie within the granite area. The slate in the band around the granite is graphitic, and contains beds of black quartzite. It is folded, dips away from the granite, and passes under the Great Smoky quartzite in normal sequence. Where seen by the writers this marginal belt of slate is not intruded by granite. The relation of the granite to the slate is confused, however, because the granite in places is sheared to a mylonite on cross fractures, which also offset the slate. Bayley (1928, p. 104) did not specifically call the slate in the inliers Hiwassee slate, but described them as "several layers of a black micaceous schist, identical with that which has been correlated with the Hiwassee slate of more northern quadrangles." Probably, the black mica schist within the granite area, which Bayley stated is saturated with granitic material, is not Hiwassee slate but is a remnant of older mica schist which was intruded and replaced by the granite, and both the injected mica schist and the Salem Church granite are parts of the injection complex which is older than the Ocoee series. The Hiwassee slate as mapped around the granite underlies the Great Smoky quartzite, and the writers interpret it as the basal formation of the Ocoee series equivalent to the Hurricane graywacke.

The Corbin granite is exposed southwest of the Salem Church granite and just east of the Cartersville overthrust, along which the Corbin granite is sheared to an augen gneiss and some layers to a sericitic or graphitic schist. Basal conglomerate beds of the Pine Log conglomerate of the east end of Pine Log Mountain, which lies north of the Corbin granite and east of the fault, are derived from the Corbin granite and rest on that granite which is brought up in an anticline (Hayes, Cartersville folio; 1901, p. 406).

The Fort Mountain (granite) gneiss (Hayes, Dalton folio) forms a narrow belt 4 to 5 miles long on the west slope of Fort and Cohutta mountains and lies south of the Holly Creek embayment of the Cartersville thrust block. Hayes states that the gneiss is overlain by a basal arkosic conglomerate, black slate, graywacke, and conglomerate. Furcron and Teague (1946, p. 1195) confirm the conclusion of Hayes that these sediments are unconformable on the granite gneiss in that area. The lower part of the sedimentary series that overlies the granite gneiss should be Hurricane graywacke.

Rocks beneath the Hurricane graywacke, exposed in anticlines.—Hayes (Cartersville folio) described the Corbin granite as ranging from massive coarse-grained augite granite to quartz gabbro. It contains blue quartz and plagioclase feldspar and has microcline phenocrysts up to 2 inches in length. The augite of the gabbro is altered to uralite. Blue quartz-microcline pegmatite cuts the Corbin granite. Bayley (1928, p. 103) states "the Salem Church granite is a white to light-gray gneissoid granite in which are dull-white and gray glistening grains of feldspar and streaks and veins of bluish quartz. . . . . In many places it is very coarse and pegmatitic." The Salem Church granite resembles phases of the Corbin granite and, because of this similarity, has been called Corbin granite (Furcron and Teague, 1945, p. 43).

The Fort Mountain gneiss (Hayes, Dalton folio) also is similar to some phases of the Corbin granite. The Fort Mountain area (Furcron and Teague, 1946, p.

1195) contains the Cohutta schist, comprising biotite-chlorite schist and dolomite, which is older than the igneous part of the Fort Mountain gneiss. The marble is in part altered to talc and serpentine along shear zones that are supplementary to the Cartersville overthrust which lies close by on the west side of the Fort Mountain anticline. The talc and serpentine (Hayes, Dalton folio; Crickmay, 1936, p. 1386) have been regarded by some as alteration products of basic igneous rocks, probably because basic rocks occur in the area of the Fort Mountain gneiss and also because serpentinized dolomite is exposed only in the newer talc mines (A. S. Furcron, personal communication, 1947).

It seems evident that the rocks called Corbin and Salem Church granites and Fort Mountain gneiss, since they include gabbro, diorite, biotite augen gneiss, granite pegmatite, and mica schist, are not a single granite intrusion or formation but are hybrid rocks formed by replacement of diorite, gabbro, or mica schist by granitic solutions. The Cohutta schist resembles the mica schist described by Bayley (1928, p. 48–49) as "inliers" in the Salem Church granite. The similar metasedimentary, igneous, and hybrid rocks in the three anticlines resemble the early pre-Cambrian injection complex. For these reasons the writers regard the rocks beneath the Hurricane graywacke in the three anticlines as part of the injection complex.

GREAT SMOKY QUARTZITE

General distribution.—The Great Smoky quartzite overlies the Hurricane graywacke and underlies the Nantahala slate. It extends from the northeast end of the belt of the Ocoee series (Roan Mountain quadrangle southwestward across northwestern North Carolina and southeastern Tennessee, passes into Georgia, and continues to a point 10 miles east of Cartersville. Because of its resistant character it forms most of the mountains and higher hills in the belt of the Ocoee series. This quartzite may be present also in the Cartersville thrust block in Alabama where quartzite, which Butts (1926, p. 52, 60) called the Cheaha sandstone member of the Talladega slate and correlated with the Great Smoky conglomerate (quartzite), forms Cheaha and Rebecca mountains and many lower hills.

At the northeast end of the belt of the Ocoee series the Great Smoky quartzite is exposed in the mountains southwest of the Limestone Cove window and between that window and the Taylor Valley window (Pl. 1). In the main part of the Great Smoky block to the west, the Great Smoky quartzite extends from the southwest side of Buffalo Mountain (Fig. 6) southwestward and surrounds the syncline of higher formations of the Ocoee series in the Bald Mountains. A section of the quartzite in the vicinity of Gravel Knob on the south side of these mountains is given (Table 1). Southwestward from the southwest end of Camp Creek Bald (Fig. 5), where the syncline rises, the quartzite forms the curving crest and slopes of the Bald Mountains as far as the Hot Springs window (Stose and Stose, 1947, Fig. 1). North of the window the quartzite lies in a belt only half a mile wide between that window and the Great Smoky overthrust. Southwestward the quartzite, in a narrow syncline in the underlying Hurricane graywacke, extends along the Bald Mountains to Snowbird Mountain, just northeast of Pigeon River. Southwest of the

river, with widening outcrop, it enters the Great Smoky Mountain National Park and makes the crest and north and south slopes of the Great Smoky Mountains.

The Great Smoky quartzite that follows the crest of the Great Smoky Mountains from Pigeon River to Newfound Gap is in the main syncline of the Ocoee series. In the vicinity of the gap (Pl. 2) this syncline deepens and encloses the Nantahala slate which the quartzite encircles. South of Newfound Gap the main syncline probably continues in the Murphy syncline which contains the Nantahala slate and younger formations with the Great Smoky quartzite on both limbs of that syncline.

In the heart of the Great Smoky Mountains west of Newfound and Indian gaps and west of the Murphy syncline, the quartzite is in a folded belt 25–30 miles wide which borders the North Carolina-Tennessee State line and passes south into Georgia. In places (Pl. 1) the quartzite contains infolded synclines of Nantahala slate. In the northern part of this folded belt, the quartzite in synclines in the Hurricane graywacke crops out in Webb Mountain (Fig. 3) and in the high mountains that surround the windows of Wear, Tuckaleechee, and Cades coves (Figs. 3, 4; Pl. 1). Northwest of these windows the quartzite is near the border of the Great Smoky overthrust block, and in Pine Mountain (Fig. 3) and other mountains south of Walden Creek, it is at the front of that block.

In northern Georgia, where Hayes (Cartersville folio; Dalton folio) called the Great Smoky quartzite the Cohutta and Pine Log conglomerates in part, it underlies Doogan, Cohutta, and Fort mountains and extends southward into the east end of Pine Log Mountain. The quartzite overlies the Hurricane graywacke (Hiwassee slate of Bayley, 1928) on the flanks of the Salem Church anticline.

South of the main crest of the Great Smoky Mountains and northeast of Cherokee (Pl. 1), the quartzite is exposed east of the Straight Creek anticline in the Balsam Mountains (Mt. Guyot quadrangle). The quartzite on the west side of the anticline extends southwestward across North Carolina, and south of Bryson in a synclinal fold it passes north into the Alarka Mountains (Pl. 1). In Georgia a belt of the Great Smoky quartzite on the southeast side of the Tusquitee-Whitestone fault and of the Murphy syncline lies between the Georgia-North Carolina State line at the northeast and Mt. Oglethorpe at the southwest.

Lithologic character.—The lower member of the Great Smoky quartzite consists of thick-bedded white to gray arkosic quartzite. The middle member is light- to dark-gray, poorly bedded quartzite containing pebble beds. The upper member is made up of coarse-grained dark quartzite, pebbly in places, and interbedded black slaty graywacke. The Great Smoky quartzite consists principally of angular grains of quartz and feldspar. The quartz is gray to blue, and glassy or milky. The feldspar is commonly white but is locally pink in the lower member. The dark color of the rock is due to the presence of fine particles of iron oxide, carbonaceous material, and in places biotite. At some localities the angular pebbles in the conglomeratic beds of the middle member comprise quartz, feldspar, and granite gneiss; at others, they are entirely vein quartz. Only the pebbles of black fine-grained, slaty graywacke, commonly with slickensided surfaces, exceed half an inch in diameter. They lithologically resemble beds in the Great Smoky quartzite and Hurricane graywacke,

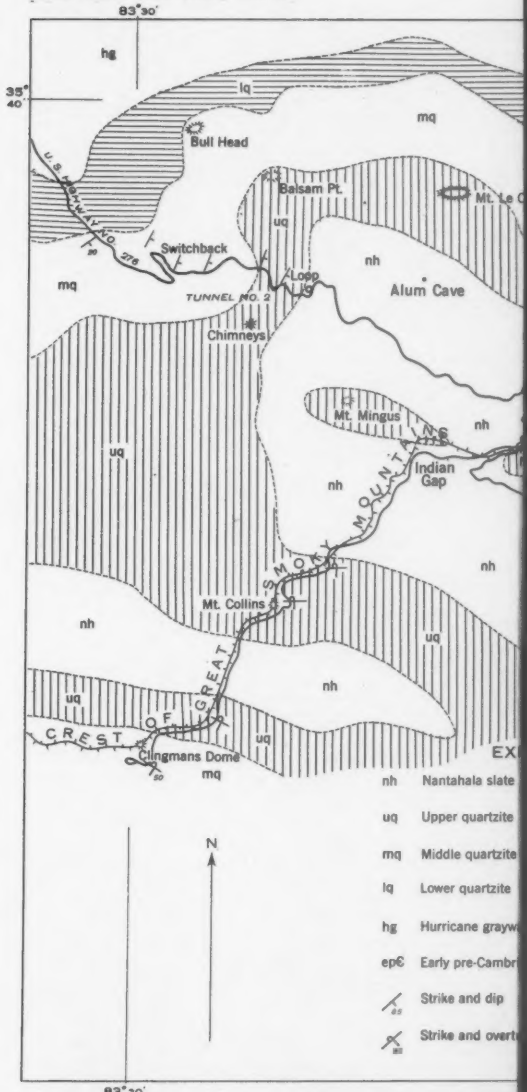
hence may be reworked material from these formations. Other detrital material is derived from metamorphosed igneous rocks similar to the injection complex.

In the section of the Ocoee series on the south side of the Big Butt, the only good exposures of the Great Smoky quartzite are of the middle (conglomeratic) member which crops out at Gravel Knob, hence the estimated thickness of 500–800 feet, given in Table 1, may be too small. The most complete and best-exposed section of the Great Smoky quartzite seen by the writers is on U. S. Highway 276 (Tennessee Highway 71) which ascends the Great Smoky Mountains southeastward to Newfound Gap (Pls. 1, 2). This highway crosses the quartzite from a point 1 mile northwest of Tunnel No. 2, where the basal beds dip 20° SE. off of the Hurricane graywacke, southeastward to the vicinity of Tunnel No. 2 (Pl. 2), where the upper beds of the quartzite dip 20° SE. under the Nantahala slate. The estimated thickness of the Great Smoky quartzite in this section is 4100± feet, and the section is as follows:

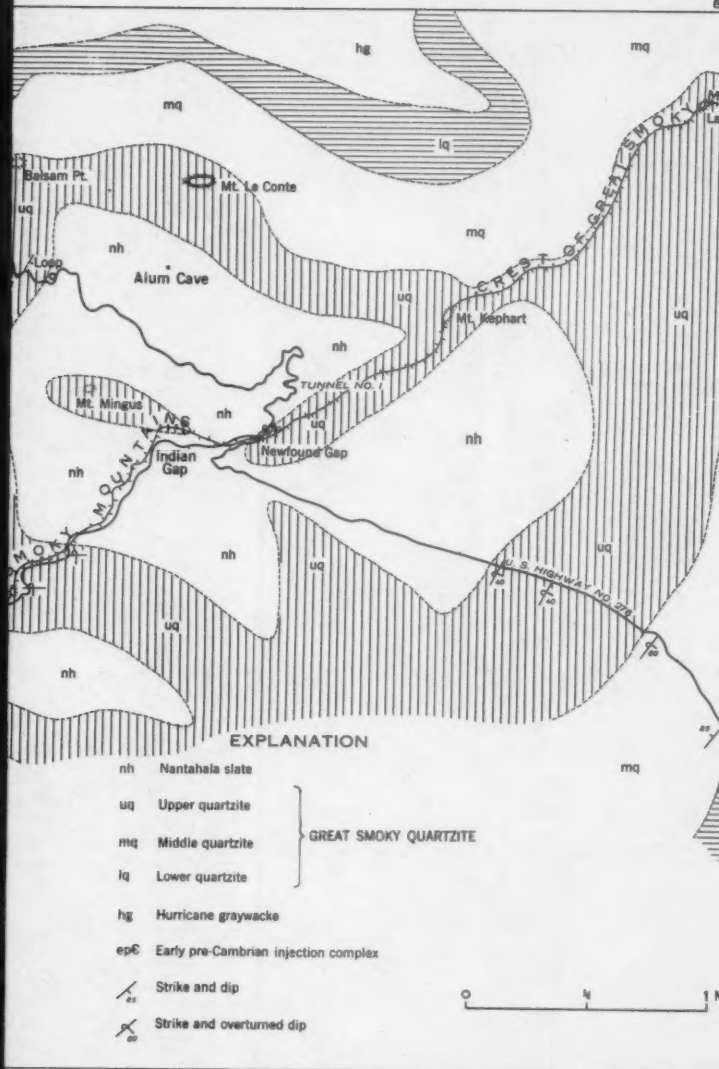
Section on U. S. Highway 276 on west slope of the Great Smoky Mountains northwest of Newfound Gap.
(Pl. 2)

	Feet
Nantahala slate (east of the Loop)	
Black shiny slate and some dark-green quartzite	
Great Smoky quartzite	
Upper member (west of the Loop to 1500 feet ± west of tunnel No. 2)	
Thick-bedded gray quartzite; upper beds, coarse dark quartzite in beds 3–5 feet thick with interbedded thin black slate; lower beds, very thick-bedded (20 foot beds) in part finely conglomeratic	1000±
Poorly exposed softer quartzose beds with some black slate	600±
Middle member (from switchback east to 1500 feet ± west of tunnel No. 2)	
Thick-bedded (20 foot beds) dark-gray quartzite, graywacke, and conglomerate containing coarse grains of angular quartz and feldspar and black fine-grained slaty graywacke pebbles	400±
Poorly exposed softer quartzose beds with some black slate	400±
Thick-bedded gray quartzite with black slate partings 1 inch thick; lower beds, very thick bedded (20-foot beds) quartzite and conglomerate of coarse blue quartz and feldspar pebbles and containing egglike concretions	600±
Lower member (west of switchback)	
Thick-bedded white and gray arkosic quartzite and fine conglomerate of white quartz pebbles	1100±
	4100±
Hurricane graywacke	
Graywacke, impure quartzite, and argillite	

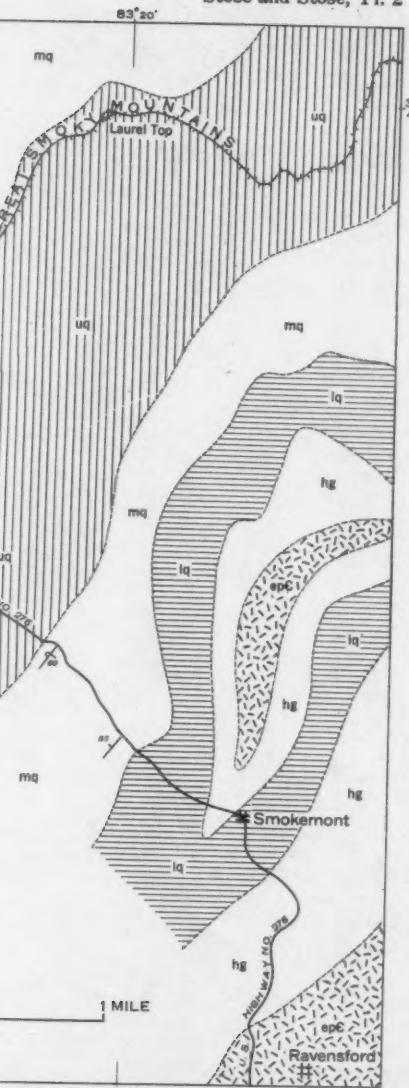
In the Knoxville and Mt. Guyot quadrangles Keith (1895b; Mt. Guyot folio) mapped three conglomerates—the Cades, Thunderhead, and Clingman, which, in the region to the south, he later classified as Great Smoky. The Cades conglomerate (Keith 1895b; Mt. Guyot folio) is the lower member of the Great Smoky quartzite (Pl. 2). The Thunderhead conglomerate forms the middle and upper members. The Nantahala slate which overlies the upper member of the Great Smoky was called the Hazel slate. The quartzite which lies east of the Nantahala slate at Newfound Gap, Keith called Clingman conglomerate, and, because the beds dip 50° SE., he



GEOLOGICAL MAP OF GREA
OF NEWFOUND



MAP OF GREAT SMOKY MOUNTAINS IN
NEWFOUND GAP, TENN. - N. C. STATE L



NS IN THE VICINITY
TATE LINE

interpreted the quartzite as overlying the Nantahala (Hazel) slate. Structural evidence shows that these beds are overturned, and that the quartzite here is the upper member of the Great Smoky on the east overturned limb of a syncline enclosing Nantahala slate (Pl. 2).

South of Newfound Gap the upper quartzite member is exposed at Mt. Collins on the Park road that leads to Clingmans Dome. Southwest of Mt. Collins, Nantahala slate is in a narrow syncline, on the south limb of which are exposed the underlying upper quartzite of the Great Smoky formation, and, farther south, the middle (conglomeratic) beds. These conglomeratic beds in the outcrops at the Parking Overlook at Clingmans Dome are massive micaceous quartzite in 20-foot beds which contain coarse pebbles of quartz and feldspar, flat pebbles of fine-grained black graywacke, and egg-shaped concretions up to 12 inches in diameter (Pl. 3, fig. 1). The concretions are composed of grains of quartz, feldspar, calcite, muscovite, and biotite and, except for a slightly darker color, resemble the enclosing quartzite. The longer diameters of the concretions are roughly parallel to the bedding, and their contact with the enclosing quartzite is sharp. They weather out of the quartzite like cobble and leave rust-stained depressions or cavities. It is these beds that Keith (Mt. Guyot folio; 1895b) called the Clingman conglomerate, which he regarded as the upper member of the Great Smoky (Table 2) and so mapped it at Clingmans Dome and on the crest of the Great Smoky Mountains to the west. The conglomeratic quartzite that crops out at Clingmans Dome is identical in lithologic character and stratigraphic position with the middle member of the Great Smoky quartzite exposed on U. S. Highway 276 and lies well down in the quartzite. The writers found nodular concretions in quartzose conglomerate of the middle member also at the border of the Great Smoky overthrust on Pine Mountain and south of Walden Creek (Citico conglomerate of Keith, 1895b), where Cove Creek cuts through that mountain (Fig 3), and at other localities.

In the type section of the Ocoee series on Ocoee River the Great Smoky quartzite crops out on U. S. Highway 64 from the sharp bend in the river at the east edge of the Cleveland quadrangle southeastward for a mile, to a point where the quartzite dips 60° SE. under the Nantahala slate. The section resembles that described northwest of Newfound Gap, but the writers did not observe the characteristic nodular concretions in the middle member. The estimated thickness of the Great Smoky quartzite on Ocoee River is 1950 feet, compared with 500–800± feet estimated to be present on the south side of the Big Butt and 4100± feet in the Great Smoky Mountains northwest of Newfound Gap. It is evident that the Great Smoky varies in thickness and is thickest in the Great Smoky Mountains.

Metamorphosed phase of the quartzite.—The Great Smoky quartzite is metamorphosed and closely folded on the southeast side of the belt of the Ocoee series from the southeastern part of the Mt. Guyot quadrangle southwestward to northern Georgia. It contains biotite, muscovite, and in places small garnets, and the intercalated slaty layers are mica schist. The pebbles in the conglomeratic member are flattened but are still recognizable, hence the middle conglomeratic member is the horizon most easily identified. The metamorphism is greatest at the southeast and wanes northwestward. High-rank metamorphism in the Great Smoky quartz-

ite is confined to the southeast side of the Great Smoky Mountains and the vicinity of the Murphy syncline. The zone of waning metamorphism is at the crest of the Great Smoky Mountains at Newfound Gap, extends southwestward and crosses Little Tennessee River (Pl. 1) about 10 miles west of the confluence with Tuckasegee River, lies about 5 miles west of Ducktown, Tenn., and in Georgia is on the west side of the Great Smoky quartzite on the west limb of the Murphy syncline.

In the highly metamorphosed Great Smoky quartzite the egg-shaped nodular concretions also are metamorphosed and have been called "pseudodiorite" (Emmons, Laney, and Keith, 1926, p. 19–21, Pl. 7C, p. 8–14; La Forge and Phalen, 1913, p. 7–8). They occur in the pebbly biotitic quartzite of the middle member in the vicinity of Ducktown and to the east. Photographs of pseudodiorite concretions (Pl. 3, fig. 2) were taken by Keith west of Murphy, N. C., in 1911 and labeled "lenticular dioritic replacement of graywacke." In his later work (Emmons, Laney, and Keith, 1926, p. 20) he recognized the sedimentary origin of the pseudodiorite and described it as follows: "Light-colored minerals—quartz and feldspar, and in places calcite and zoisite—form the greater part of the rock and give it a distinctly light-gray color. Hornblende, garnet, and biotite occur as metacrysts." The texture of the pseudodiorite in thin section and the relative abundance of the constituent minerals—the presence of 25 per cent calcite and abundant zoisite with no indication of replacement to account for their formation—are given as evidence that the pseudodiorite was derived from a sedimentary rock. Bayley (1928) described a thin section of pseudodiorite from the Great Smoky quartzite on the east side of the Salem Church anticline in Georgia as consisting "mainly of interlocking quartz, (cellular) garnet, pale green hornblende, and epidote, all strung out in lines and in some (thin) sections there are also a little plagioclase, sphene, zircon, apatite, and magnetite."

In northern Georgia the belt of Great Smoky quartzite east of the Murphy syncline and of the Whitestone-Tusquitee thrust fault was mapped (La Forge and Phalen, 1913) only in a narrow band along the east side of the thrust fault and in an isolated area to the southeast. La Forge and Phalen included the remainder of the quartzite there exposed in the Carolina gneiss. A pebbly quartzite of the middle member, which forms a ridge just east of the Whitestone fault at Ellijay, represents all the Great Smoky formation there mapped. To the southeast, beds typical of the Great Smoky crop out at several places, and, 4 miles southeast of the fault, pebbly biotitic quartzite containing pseudodiorite nodules characteristic of the middle member of the Great Smoky is exposed in a quarry; hence it should not be called Carolina gneiss. The Great Smoky quartzite extends south of Ellijay to Mt. Oglethorpe, at the south end of the Blue Ridge, where it has been named the Oglethorpe formation (Furcron and Teague, 1945, p. 35–36, Fig. 2, Pl. 3). Near Mt. Oglethorpe it is a biotitic quartzite containing pseudodiorite in nodular concretions and also in cylindrical bodies that are similar to the "dike-like" bodies (Emmons, Laney, and Keith, 1926, p. 20) that occur near Ducktown and in outcrops seen by the writers west of Robbinsville, N. C. (Pl. 1).

NANTAHALA SLATE

The Nantahala slate is dense, bluish-black, shiny, and graphitic, and commonly contains much pyrite. The writers use the term Nantahala slate as defined (Keith,

FIGURE 1. EGG-SHAPED NODULE
At Clingmans Dome in middle member of formation.

FIGURE 2. PSEUDODIORITE NODULE, WEST OF MURPHY
Photograph by A. Keith

EGG-SHAPED NODULES IN THE GREAT SMOKY QUARTZITE

FIGURE 1. ROUND QUARTZ-PEBBLE CONGLOMERATE
At base of Unicoi formation. Green Cove Creek Gorge, Va.

FIGURE 2. RIPPLE MARKS IN TUFF OF MT. ROGERS VOLCANIC SERIES
Evidence of little metamorphism of late pre-Cambrian sediments. Green Cove, Va.

CONGLOMERATE AND RIPPLE MARKS

1907a, p. 4) in the type locality on Nantahala River, N. C., where Keith mapped the slate as overlying the Great Smoky conglomerate (quartzite).

The Nantahala slate is present above the Great Smoky quartzite in the main syncline of the Ocoee series in the Bald Mountains, in the Great Smoky Mountains in the vicinity of Newfound Gap and southwestward, and on the flanks of the Murphy syncline from the northeast corner of the Nantahala quadrangle southwestward into northern Georgia. In the deeper parts of the syncline, as in the Bald Mountains and in the Murphy syncline, the Nantahala is overlain by the Big Butt (Tusquitee) quartzite. Northwest of the main syncline, from the vicinity of Little Tennessee River southward across southern Tennessee and North Carolina, the Nantahala slate is enclosed in minor synclines in the Great Smoky quartzite (Pl. 1).

The Nantahala slate near the top of the Bald Mountains is described in Table 1. Because the slate there is poorly exposed, the estimated thickness of 200 feet is approximate. Near Newfound Gap (Pl. 2) in the Great Smoky National Park, the slate lies in the broad syncline of Great Smoky quartzite that northeastward follows the crest of the Great Smoky Mountains. In this syncline the slate crops out on the crest of the mountain at the west side of Newfound Gap, at Indian Gap, and southwestward to the vicinity of Mt. Collins; also in a narrow syncline between Mt. Collins and Clingmans Dome. On the lower north slope of the mountain, massive gray quartzite of the Great Smoky dips 20° SE. under the Nantahala slate in the main syncline. Southeastward to Newfound Gap the slate is closely folded, and the folds are overturned to the northwest.

The Nantahala slate in these synclines is a black, graphitic, shiny slate typical of the Nantahala and has thin beds of black quartzite and thicker beds of rusty-weathering gray quartzite. Pyrite, where present, oxidizes to a rusty stain and in some places to sulphates forming the so-called "alum slates" which exhibit an aluminous efflorescence. Alum Cave (Pl. 2) takes its name from this characteristic weathering. A partial section of the Nantahala slate in exposures on the west side of the Parking Overlook at Newfound Gap (Pl. 2) is as follows:

Partial section of Nantahala slate, on U. S. Highway 276, Newfound Gap

	Feet
Nantahala slate	
Black slate	
Grayish green quartzite or graywacke	20
Slabby quartzite and interbedded black slate	25
Grayish green quartzite	6
Black slate	20
Thin grayish green quartzite	2
Black graphitic slate containing small garnets	80±
Great Smoky quartzite	
Coarse gray arkosic quartzite and conglomerate with black slaty graywacke pebbles and lenticular layers	10
Thick-bedded hard gray quartzite	60±

The Great Smoky quartzite crops out in the knoll at the east side of the Parking Overlook at the gap, and the contact with the Nantahala slate is exposed at the north end of the road cut on the west side of the highway. The beds dip 50° SE., overturned.

Southwest of Newfound Gap the Nantahala slate is present on the limbs of the Murphy syncline in linear curving bands across North Carolina into Georgia. In this area the slate overlies the Great Smoky quartzite and is overlain by the Tusquitee quartzite (Keith, 1907a). At the northeast end of the Murphy syncline, where the Tuckasegee River cuts through the Nantahala slate for about 5 miles, the slate is metamorphosed to a black, graphitic, ferruginous, quartzose mica schist studded with garnet, staurolite, and ottrelite. To the southwest near Nantahala River, the metamorphic minerals in the Nantahala are smaller and less abundant, and farther southwest are not present. Southwest of Mt. Oglethorpe, hard ridge-making dark quartzite with interbedded black slate containing much pyrite forms the lower part of the Nantahala and is overlain by black graphitic pyritiferous slate. The formation here is in close folds overturned to the northwest.

The Nantahala slate in the type section on Ocoee River (Safford, 1856) is in a syncline in the Great Smoky quartzite which dips under the slate on both sides of the fold, and only the lower beds of the formation, represented by black slate and thin black quartzite, are exposed.

BIG BUTT QUARTZITE

The Big Butt quartzite forms the top of Big Butt and adjoining high peaks to the southwest to the vicinity of Camp Creek Bald (Fig. 5) on the crest of the Bald Mountains. At many points along this mountain top the quartzite crops out in prominent white cliffs which can be seen for many miles from either side of the range. On the Big Butt the formation consists of two beds of massive white coarse-grained quartzite, each about 30 feet thick, separated by approximately 100 feet of dark-green, fine-grained, dense arkose and argillite. Northeast of the Big Butt the quartzite caps Rich Mountain (Roan Mountain quadrangle) and in the vicinity of Nolichucky River forms Jump Hill and caps Stony Point. At all these places the quartzite is enclosed in the main syncline of the Ocoee series and overlies the Nantahala slate. The Big Butt quartzite is not present in the main syncline that follows the crest of the Great Smoky Mountains southwest of Camp Creek Bald to Newfound Gap, where the deeper parts of the syncline do not contain beds younger than the Nantahala slate, and the quartzite is not known in the Great Smoky Mountains to the southwest.

In the Murphy syncline, Keith (1907a) and La Forge and Phalen (1913) mapped a white quartzite—the Tusquitee—which overlies the Nantahala slate and underlies the Brasstown schist and Valleytown formation. Keith describes the Tusquitee quartzite as a white quartzite 50 to 200 feet thick occurring in beds 3 feet thick or less. At most places he mapped it in two parallel lenticular bands with anticlines of Nantahala slate between. The writers correlate these white quartzites with the Big Butt, and they regard the intervening slaty beds not as Nantahala slate but as part of the Big Butt quartzite equivalent to the fine-grained arkosic quartzite and argillite that separate the two quartzite beds of the Big Butt formation on the Bald Mountains.

On the northwest limb of the Murphy syncline the Tusquitee quartzite as mapped by Keith (1907a) forms Cheoah Bald and the eastern part of Snowbird Mountain,

N. C. To the southwest a branch fault cuts off the Tusquitee quartzite so that at Murphy and southwestward it is not present on the northwest limb of the syncline. On the southeast limb, near the northeast end of the Murphy syncline where Nantahala River crosses the Big Butt (Tusquitee) quartzite, it appears to be in a minor syncline, and only the lower quartzite is present. The section there is as follows:

Partial section of Tusquitee (Big Butt) quartzite west of Wesser

	Feet
Tusquitee quartzite	
Schist and banded dark quartzite	?
Thin-bedded to platy quartzite	20
Thick white quartzite	5
Thin-bedded quartzite	15
Massive white quartzite	100±
	140±
Nantahala slate	
Black slaty quartzite and black-banded quartz schist	

One mile southeast of Murphy (Pl. 1) the Big Butt quartzite makes a prominent ridge which is cut through by the Hiwassee River. In exposures on the north side of the river, this quartzite is 200 feet thick and consists of two layers of massive thick-bedded white quartzite with thinner beds of dark-banded, white and blue, irregularly bedded quartzite in the middle.

In the northern part of the Ellijay quadrangle the quartzite is apparently 500 to 600 feet thick (La Forge and Phalen, 1913), but it thins southwestward and is not known south of Ellijay where the Valleytown formation lies directly on the Nantahala slate.

VALLEYTOWN FORMATION

Green phyllite, slate, and ottrelite-mica schist overlie the Big Butt (Tusquitee) quartzite in the Murphy syncline. In the type locality in the Nantahala quadrangle, Keith (1907a, p. 4) separated these rocks into the Brasstown schist and Valleytown formation but states that "the mica schist (of the Valleytown) passes downward into the Brasstown schist, and individual layers of each formation cannot be distinguished from those of the other". Only locally have the writers found lithologic criteria for separating these rocks into two formations. Furthermore these phyllites and schists for the most part are closely folded and have a prominently developed schistosity which obscures the bedding, so that, on stratigraphic evidence available, they are not readily divisible. In the Ellijay quadrangle La Forge and Phalen (1913) recognized the Brasstown schist only in the northern part of the quadrangle; and in Georgia, Bayley (1928, p. 65–73) included these rocks in a single formation for which he used the name Valleytown. These two formations, therefore, are here treated as Valleytown formation, which is the more widely used name.

The Valleytown formation is composed of argillaceous and siliceous rocks which have been metamorphosed to green phyllite, mica schist, and garnetiferous mica schist, which in places contains ottrelite. In the vicinity of thrust faults, the ottrelite

may be a retrogressive alteration of biotite. In places the mica schist contains staurolite and kyanite crystals. As here used the Valleytown formation, which is not known outside the Murphy syncline, extends from the northern part of the Nantahala quadrangle southwestward across the Murphy quadrangle into Georgia, trends south across the Ellijay and Tate quadrangles, enters the Cartersville quadrangle, and is cut off at the south by the Hillabee overthrust about 6 miles southwest of Canton.

South of Ellijay, where the Big Butt (Tusquitee) quartzite is absent, the Valleytown formation overlies the Nantahala slate. The Valleytown formation is overlain by the Nottely quartzite and Murphy marble, but the quartzite is not everywhere present, and in places (Fig. 1) the Murphy marble lies directly on the Valleytown formation.

CAMBRIAN ? SYSTEM

MURPHY MARBLE AND NOTTELY QUARTZITE

Distribution.—The Murphy marble is present only in the Murphy syncline. It is discontinuously exposed for nearly 100 miles from the vicinity of Nantahala River, 2 miles northeast of Hewitts, southward to Canton, Ga., where the syncline and the Murphy marble are cut off by the Hillabee thrust. The longest continuous belt of marble extends for 30 miles along the main part of the syncline from Valleytown, N. C., southwestward through Murphy into Georgia (Pl. 1). Southwest of Mineral Bluff (Fig. 2) the marble is exposed in disconnected southwest-trending areas. Just east of Tate (Fig. 1) it crops out in the east-west valley of East Branch Long Swamp Creek where it is extensively quarried for commercial ("Tate") marble (Bayley 1928, p. 146–147). In southeastern North Carolina, southeast of the main belt, disconnected outcrops of Murphy marble in a minor syncline pass through Peachtree, N. C. and extend southwest into Georgia to a point 1 mile east of Blue Ridge (Fig. 2). In Plate 1 the writers have included with the Murphy marble the calcareous Andrews schist which is associated with the marble in the northeastern parts of both synclines.

The Nottely quartzite is a thin but prominent ridge-making quartzite which occurs in contact with the Murphy marble only in the main belt from Ellijay, Ga., northeastward. From Mineral Bluff, Ga., northeastward to Tomotla, N. C. (Fig. 2), it is exposed continuously for more than 24 miles. The quartzite dips steeply southeast to vertical. In places it is folded with the marble, but for the most part it lies southeast of the marble and northwest of the Valleytown formation. In the Nantahala quadrangle (Keith, 1907a, p. 5) the quartzite is 150 feet thick, and in the Ellijay quadrangle (La Forge and Phalen, 1913, p. 7) it is 200 feet thick. Where the Nottely quartzite is not present and the Murphy marble lies adjacent to the Valleytown formation, as in the Tate area, Ga. (Fig. 1), the writers assume that the Nottely is absent because of nondeposition.

Structural relations.—The Murphy syncline is a tightly compressed, compound syncline, and at most places is bounded by thrust faults. These faults cut off the Murphy marble at the northeast, and at many places cut out both the marble and Nottely quartzite. In places in the northeastern part of the syncline the Valley-

town formation is on both sides of the marble, apparently in fault contact. Elsewhere the marble, and in places the Nottely quartzite, are in fault contact with the Nantahala slate and Great Smoky quartzite. These faults, as well as the variable depth of the syncline, account for the discontinuity of outcrops of the Murphy marble and, in the northern part of the syncline, of the Nottely quartzite.

The trend of the syncline and of the formations in the fold is southwest as far as Whitestone (La Forge and Phalen, 1913, p. 9, figs. 5, and 6). There the Whitestone thrust, and the Murphy marble which is exposed in discontinuous outcrops west of the fault, bend east and extend to the northern part of the Tate quadrangle, where they again trend southwest. In the northern part of the Tate quadrangle the thrusts along the Murphy syncline, including the Whitestone fault, dip about 45° SE. and are straight or gently curving. A mile northeast of Tate station the Whitestone fault is offset 1 mile to the northeast by a diagonal normal fault (Fig. 1). South of this normal fault the Whitestone thrust has a sinuous trend and swings eastward 4 miles up the valley of East Branch Long Swamp Creek and around the Marble Hill area of the "Tate" marble. The marble in this valley is bordered and overlain by gneiss (Carolina gneiss of Bayley) which is thrust over the marble. The east-west trend of the area of marble and the sinuous trace of the fault, which are so markedly at variance with the linear southwest trend of the marble and the fault elsewhere, are explained by Bayley (1928, p. 123–126, Fig. 2) as follows: "Here the fault plane and rocks on both sides have been compressed into a major east-west syncline with subordinate cross folds." The sole of the thrust in the Marble Hill area and the layering of the marble beneath the thrust are almost horizontal, and the marble probably underlies the gneiss both north and south of the valley of East Branch. The writers regard the sinuous trace of the fault plane as the result of the erosion of a flat overthrust plate and attribute the long east-trending exposure of the marble as due to the horizontal attitude of the overridden marble in a broad syncline.

At Mt. Oglethorpe (Pl. 1), about 3 miles north of the Marble Hill area, the gneiss is overlain by the Hurricane graywacke and Great Smoky quartzite at the south end of the Mt. Oglethorpe syncline. The gneiss has been called the Amicolola gneiss (Furcron and Teague, 1945, p. 31). Because of the stratigraphic position and lithologic character of the gneiss, the writers regard it as part of the early pre-Cambrian injection complex.

The rocks of the Ocoee series in the Mt. Oglethorpe syncline, which rode westward on the Whitestone thrust over the center of the Murphy syncline, are terminated by the overthrust just north of the normal fault northeast of Tate, and farther south the underlying Amicolola gneiss is at the sole of the Whitestone overthrust. Just south of Tate (Fig. 1) the gneiss overrides and conceals the Murphy marble, but it appears again just west of the Whitestone fault at a point east of Nelson and extends southward through Ballground. Northeast of Ballground the presence of a narrow band of southeast-dipping marble within the overthrust gneiss suggests that the fault plane here is sharply folded.

West of Ballground narrow areas of marble crop out in several synclines in the Valleytown formation (Fig. 1). The central syncline, which passes through Nelson and east of Tate station, is offset from its northward continuation in the valley of Long Swamp Creek by the normal fault.

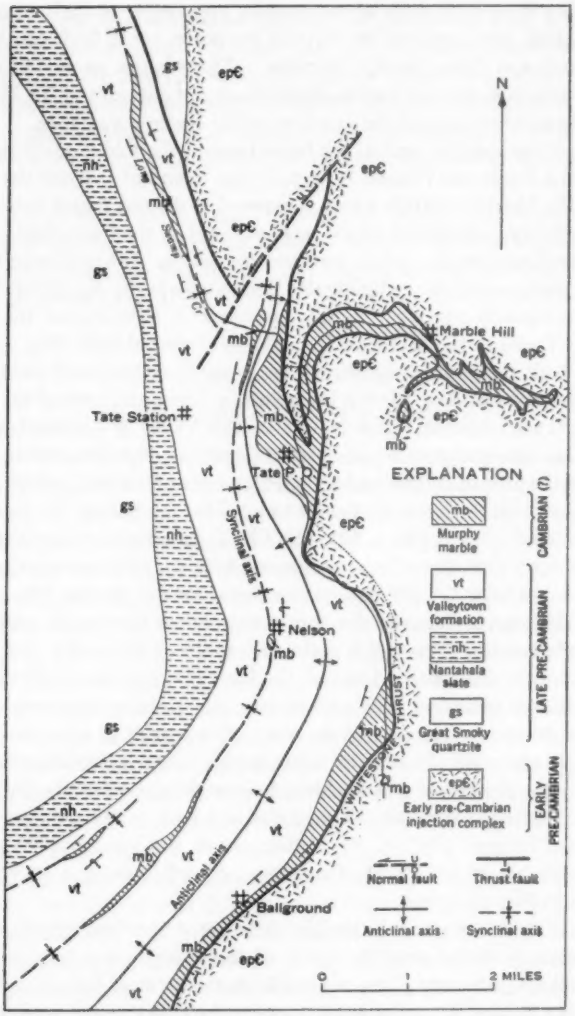


FIGURE 1.—*Geological map of part of the Tate quadrangle, Georgia*

By W. S. Bayley, with structural interpretation by the writers. The Hurricane graywacke has not been separated from the Great Smoky quartzite.

Sequence of formations.—All geologists who have worked in the region agree that the Murphy marble, Andrews schist, and Nottely quartzite lie above the Valleytown formation. According to Keith (1907a) the succession of formations, in rising sequence, is Valleytown formation, Murphy marble, Andrews schist, and Nottely quartzite. The only locality where the writers have found the Valleytown formation

apparently in normal contact with overlying formations is in the main syncline in northern Georgia where the Nottely quartzite dips 70° SE. to vertical and lies between the Murphy marble at the west and the Valleytown formation at the east. A coarse conglomerate on the east side of the quartzite is in contact with the Valleytown and has been traced by Furcron and Teague from the Georgia-North Carolina State line southwest to Coles Crossing. On State Highway 60, northeast of Coles Crossing (Fig. 2C), the conglomerate contains pebbles of vein quartz up to 3 inches in diameter. Smaller pebbles include fine-grained granite gneiss cut by quartz vein, fine-grained quartzite with a sugary texture, vein quartz containing tourmaline, and finely crystalline magnetite. These pebbles were derived from the erosion of crystalline rocks, and the pebbles of fine-grained quartzite may have been derived from the erosion of the Ocoee series. The conglomerate suggests a coarse initial sediment at the base of clastic deposits. If the conglomerate is at the base of the Nottely, that quartzite is overturned and overlies the Valleytown formation and underlies the Murphy marble at the west, and the stratigraphic order of conglomerate, quartzite, limestone is the common order of sedimentary deposition.

Structural relations at Mineral Bluff and Blue Ridge, and near Tomotla, show that the Murphy marble and the Nottely quartzite are folded together. Just west of Mineral Bluff (Fig. 2B) a hill of Nottely quartzite projects south into the marble belt, and the marble passes around the quartzite in a plunging fold. At Blue Ridge (Fig. 2A), the Nottely quartzite with conglomerate at its east margin at one place lies east of a strike valley which exposes clay derived from weathered marble. The quartzite is in a plunging fold, and the marble surrounds its north end. A prominent hill of Nottely quartzite, which extends southwestward from a point just south of Tomotla through Regal Station, lies between two belts of marble (Fig. 2D). Here again the marble passes around the north end of the quartzite, suggesting a plunging fold. The section of the Nottely quartzite exposed at that place is as follows:

Section of Nottely quartzite at Regal Station, N. C.

	Feet
Marble and dolomite	
Ottrelite schist with iron oxide deposits (Andrews schist of Keith)	
Nottely quartzite	
Crumbly ferruginous quartzite	10
Vitreous blue to white quartzite	35
Vitreous gray quartzite, weathers rusty	20
Crumbly quartzite	10
Spangled ottrelite schist (Andrews schist of Keith)	

If the succession of beds, in rising sequence, is conglomerate, quartzite, and marble, as the writers have suggested, the plunging folds illustrated in Figure 2 are anticlines that bring up the Nottely quartzite beneath the Murphy marble. Because of poor exposures the direction of pitch at the ends of the folds could not be determined. In the fold southwest of Tomotla (Fig. 2D), ledges of quartzite at the north end of the hill appear to plunge north under the marble and are not canoe-shaped as they should be if they were in a south-plunging syncline. The fault on the west margin

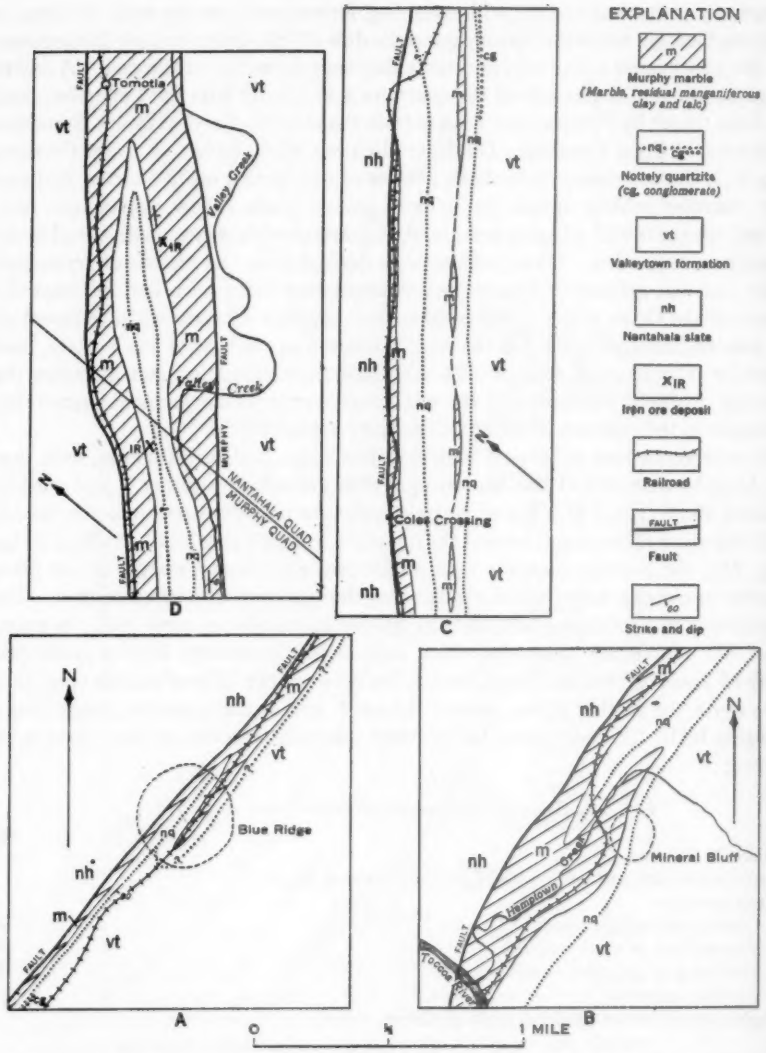


FIGURE 2.—*Geological maps showing folding of the Murphy marble and Nottely quartzite*

A. At Blue Ridge, Ga. B. At Mineral Bluff, Ga. C. In vicinity of Coles Crossing, Ga. D. In vicinity of Tomotla, N. C.

of the western band of marble accounts for the absence of quartzite at that place where it should occur if the marble is in a syncline. The medial ridge of Nottely quartzite at Coles Crossing (Fig. 2C) ends west of Mineral Bluff (Fig. 2B) and is probably a southward-plunging anticline. The Young Stone Creek fault (La Forge

and Phalen, 1913, p. 9, Fig. 5), on the west side of this folded belt of marble and quartizite, cuts off the west belt of marble and the Nottely quartzite northeast of Coles Crossing.

The Andrews schist in the Nantahala quadrangle (Keith, 1907a, p. 9–10) and Murphy quadrangle (Bayley, 1922, p. 158–163, 166–204) contains brown hematite, which occurs as layers in the weathered schist and as lumps and masses in residual clay. Bayley (1922, p. 163, 169–171) stated that the deposits are at the contact of the Nottely quartzite and Andrews schist and also are concentrated along faults in the vicinity of these formations; that the iron ores at many places are highly manganiferous, especially between Tomotla and Culberson, N. C. (Bayley, 1922, Pl. IV), where the Nottely quartzite, Andrews schist, and Murphy marble are exposed in a continuous belt. Southwestward in Georgia the Andrews schist (La Forge and Phalen, 1913, p. 7) is represented by a calcareous schist at the contact of the marble and quartzite and contains iron ore which in places is manganiferous. The writers interpret the calcareous Andrews schist as an impure transition zone between the Nottely quartzite and the overlying Murphy marble.

Age.—The Nottely quartzite and Murphy marble are present only in the deepest syncline in the Ocoee series. They were previously regarded as part of the Ocoee series and like that series were classed as Cambrian (Keith, 1907a; La Forge and Phalen, 1913; Bayley, 1928). The writers give their reasons for regarding the Nottely quartzite and Murphy marble as Cambrian. The Nottely is a clean white to blue quartzite unlike those in the Ocoee series but resembling quartzites of established Lower Cambrian age. Wherever the uppermost Lower Cambrian quartzite (Erwin, Hesse, or Antietam) is present in the Great Valley (Stose, 1919; 1942) and in the Piedmont Plateau, the top beds contain iron and manganese oxide deposits and grade into the overlying dolomite (Shady, Tomstown, or Vintage). Impure calcareous beds (Andrews schist), containing manganiferous iron-ore deposits, form a similar gradation zone between the Nottely quartzite and the Murphy marble. Limestones comparable with the Murphy marble are not present elsewhere in the Ocoee series. The writers suggest, therefore, that the Murphy marble may represent the Lower Cambrian dolomite horizon, and the Nottely quartzite the underlying Lower Cambrian quartzite; that the Andrews schist represents the iron- and manganese-bearing transition zone commonly present at the top of the Lower Cambrian quartzite; and that the conglomerate is at the base of the Lower Cambrian formations in the Murphy syncline and there overlies the youngest formation of the Ocoee series—the Valleytown formation.

TELLICO FORMATION, ERRONEOUSLY INCLUDED IN THE OCOEE SERIES

INTRODUCTORY STATEMENT

In the Mt. Guyot quadrangle and southwestward to Georgia Keith (1895b; 1896; Mt. Guyot folio) and Hayes (1895), who conformed to Keith's views, mapped the northwest side of the belt of the Ocoee series as the Wilhite slate and Citico conglomerate. The writers (1944, Fig. 2B) formerly accepted this interpretation and assumed that the thrust fault which Keith (Mt. Guyot folio) mapped on the northwest side of his Wilhite slate was the Great Smoky overthrust. The writers have since

separated the areas mapped as Wilhite slate and Citico conglomerate into two formations—the Hurricane graywacke that lies southeast of the Great Smoky thrust as now mapped (Pl. 1) and rocks of very different character that lie northwest of that thrust. These latter comprise shale, limestone, limestone conglomerate, and sandstone which do not have the dense, indurated character, dark color, and prominent cleavage of the Hurricane graywacke or other formations of the Ocoee series, but are soft, well-bedded, unmetamorphosed sediments with poorly developed slaty cleavage. The presence of Orodovician fossils, which the writers found at a few places in these beds, proves that they do not belong in the Ocoee series but are a previously unrecognized part of the Ordovician system. As they comprise Ordovician sediments of undetermined stratigraphic range in part equivalent to the Tellico sandstone (Keith, 1895b), they are here referred to as the Tellico formation rather than the Ordovician ? series (Stose and Stose, 1946b). The term Tellico sandstone will here be used as originally defined for a definite stratigraphic unit.

DISTRIBUTION

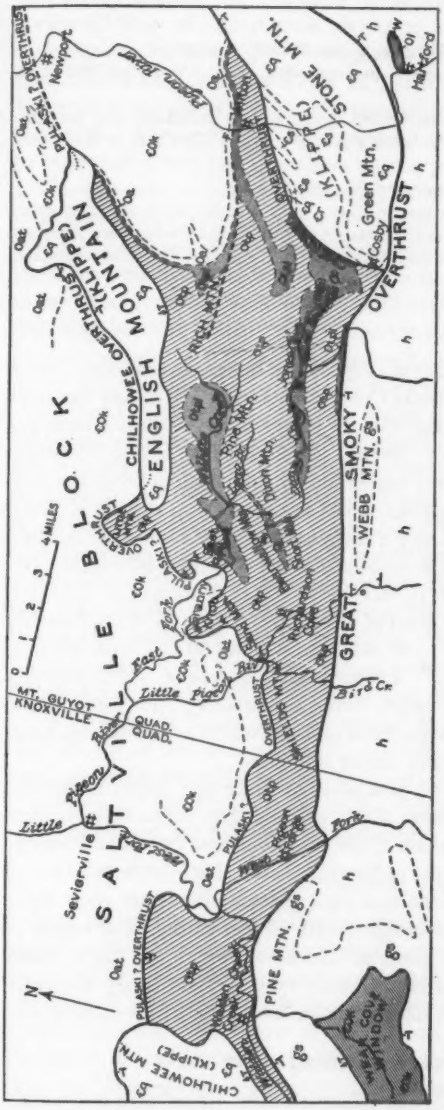
The Tellico formation occurs in Tennessee from the vicinity of English Mountain southwest to Ocoee River (Pl. 1). The largest area (Figs. 3, 4) having a maximum width of 6 miles extends for 33 miles from near English Mountain sothwest to the northeast end of Chilhowee Mountain. The formation is exposed also in narrow areas on the southeast side of Chilhowee Mountain and southeast of Starr and Beans mountains (Figs. 4, 7). The Great Smoky overthrust overrides the southeast side of the Tellico formation and overlaps it entirely in southern Tennessee and northern Georgia. Lower Cambrian quartzites of the Chilhowee group in English, Chilhowee, Starr, and Beans mountains overlie the northwest side of the Tellico formation in thrust relations.

The Tellico formation is exposed also in small areas within the Ocoee series in the Great Smoky and equivalent Cartersville overthrust block at points near Allen Stand, N. C., Delrio and Hartford, Tenn. (Figs. 3, 5), and in Alaculsy Valley, Ga. (Pl. 1). Near Parksville (Fig. 7) a small area of Tellico formation is surrounded by quartzites of the Chilhowee group.

SEQUENCE OF BEDS

On the southeast side of English Mountain (Fig. 3) Keith (Mt Guyot folio) mapped narrow bands of Athens shale and Tellico sandstone above the Knox dolomite, on its southwest border. He placed a thrust fault between the band of Tellico-Athens and the rocks to the southwest, which he mapped as Wilhite slate (of the Ocoee series). The rocks which he called Whilhite slate are soft shale, sandstone, and limestone, and overlie in normal sequence the beds in the narrow band that he mapped correctly as Tellico sandstone. The only fault present is a minor one (Fig. 3). The Knox dolomite in this area is in a southwest-pitching anticline and is overlain by fossiliferous black Athens shale followed by the Tellico formation comprising shale and limestone (the Tellico sandstone of Keith) and overlying sandstone and shale.

The general sequence of rocks of the Tellico formation in the region southwest of English Mountain is as follows:



Great Valley sequence
Oat, Athens shale and Tellico sandstone
Oa, Athens shale
€Ok, Knox dolomite (in part in windows)
Pulaski overthrust block
Otp, Tellico formation, (includes locally dolomite and limestone, 01)
Otpl, Limestone and shale at base of Tellico formation
Chilhowee and English mountains thrust blocks (klippen)
€g, Lower Cambrian quartzite *Great Smoky thrust block*
gs, Great Smoky quartzite) } Ocoee series
h, Hurricane graywacke }

T, Overthrust side of thrust fault
T↑, Direction of dip of thrust fault on east side of klippe and west side of window

FIGURE 3.—*Geological map of area between English Mountain and Chilhowee Mountain*

Showing the Tellico formation in the Pulaski block overridden by the Ocoee series in the Great Smoky block and overlain by Lower Cambrian quartzites in the English Mountain and Chilhowee Mountain klippen; Tellico formation also exposed in windows in the Great Smoky block. Adjoins Figure 4 on north.

Sequence of the Tellico formation

Tellico formation
- Hard white quartzite (exposed in Shell, Rich, Sand, and Shields mountains).
- Thick-bedded pebbly arkosic quartzite (exposed in Bearwallow, Short, Dixon, and Pine mountains).
- Calcareous quartzite, white limestone, thin-bedded dolomite, blue limestone, and limestone conglomerate which contains manganese carbonate at a point 1½ miles east of Richardson Cove.
- Thick, well-bedded sandstone and gray shale and thin purple shale.
- Blue calcareous shale, edgewise conglomerate in places, dark granular limestone with glassy quartz grains, with limestone conglomerate the matrix of which is like the enclosing limestone and the pebbles comprise fine-grained gray limestone and granular limestone containing quartz grains like the matrix.

Athens shale (in places)
Knox dolomite

The oldest beds of the Tellico formation (Fig. 3) are limestone and shale which crop out west of Cosby in the valley of Dunn Creek and its tributary Yellow Breeches Creek, on East Fork Little Pigeon River, and on its tributary, Wilhite Creek, the type locality of the Wilhite slate (Keith, 1895b). These beds are overlain by sandstone and shale, which pass upward into a higher horizon of limestones that has been recognized only in the area along Long Branch and southwestward toward Richardson Cove. Half a mile southwest of Long Branch and 2½ miles east of Richardson Cove this limestone contains manganese carbonate which has been mined for manganese ore (Stose and Schrader, 1923, p. 102–105). In that report (p. 103) it is stated that the ore follows a vertical contact of dolomite and slate "and the dolomite is apparently interbedded in the slate. The dolomite is about 60 feet thick and contains some thin-bedded blue limestone." Pebbly arkosic quartzite, which probably overlies this dolomite and associated rocks, forms Bearwallow, Short, and Dixon mountains in deeper synclines. Younger beds of hard white quartzite cap Shell, Sand, Shields, and Rich mountains in still deeper synclines.

In the areas of the Tellico formation southeast of Starr and Beans mountains (Fig. 7) gray shale and sandstone are more widely exposed than the underlying limestones of the formation. The best section of the limestone and limestone conglomerate in that belt is in a quarry northeast of Servilla, Tenn. (Fig. 7) where 30 feet of dense fine-grained bluish-black limestone and dark granular limestone with glassy quartz grains are capped by limestone conglomerate 10 feet thick. The matrix of the conglomerate is dense dark-blue granular limestone containing round glassy quartz grains, which encloses pebbles of limestone ranging from finely laminated pure, dove, fine-grained rock to dense dark granular limestone with quartz grains.

LIMESTONE IN WINDOWS IN OVERTHRUST OCOEE SERIES

In the Asheville folio (Keith, 1904), the Hiwassee slate (Hurricane graywacke) that occurs within the belt of the Ocoee series south of the Great Smoky overthrust (Pl. 1) is described as a dark banded slate and schist, with layers of limestone that are separately mapped. The limestone crops out southeast of Allen Stand and in two areas about 2 miles southeast of Delrio (Fig. 5). At all these places it is a massive

dark granular limestone with glassy quartz grains and limestone conglomerate with associated blue calcareous shale, similar to the lower calcareous beds of the Tellico formation in the main belt. Southeast of Allen Stand (Fig. 5) the limestone is in an area 3 miles long trending N. 70° E. It is surrounded and overlain by Hurricane graywacke and Great Smoky quartzite whose trend is discordant with that of the limestone. The contact of the limestone and the overlying formations is marked by masses of pitted vein quartz containing pyrite and by seeps of chalybeate water. At one place the quartz at the contact has been prospected for pyrite. The structural discordance and the presence of mineralized vein quartz at the contact indicate that the Hurricane graywacke and Great Smoky quartzite have been thrust over the limestone and that it is not part of the sequence of the Hurricane graywacke (Hiwasse slate). In the area southeast of Delrio (Fig. 5) the Hurricane graywacke (Hiwassee slate of Keith) that overlies the limestone is a sheared contorted argillite resembling deformed Hurricane graywacke elsewhere on thrust faults, and is evidently here also in thrust relation to the limestone.

At Hartford (Fig. 3) the writers found massive gray dolomite and argillaceous limestone, surrounded by Hurricane graywacke, in an area previously mapped as Wilhite slate (Keith, Mt. Guyot folio). The dolomite exposed in a quarry is veined with calcite and is slickensided along joints. The overlying Hurricane graywacke in the quarry is closely folded, crushed, and rust-stained, indicating a fault contact with the dolomite. The dolomite resembles the higher calcareous horizon of the Tellico formation. Blue limestone and overlying shale and sandstone of the lower part of the Tellico formation crop out in Alaculsy Valley, Ga. (Pl. 1), and are overlain on their northwest side by quartzites of the Chilhowee group in thrust relation. These two diverse groups of rocks are surrounded by Hurricane graywacke and Great Smoky quartzite which are thrust over both of these groups.

The limestones, shales, sandstones, and dolomite just described are stratigraphically not part of the rocks of the Ocoee series (Hiwassee and Wilhite slates), as they have previously been classed, but are part of the Tellico formation which has been overridden by the Ocoee series in the Great Smoky and equivalent Cartersville thrust block and are now exposed in windows in that block.

STRUCTURAL RELATIONS

From the southwest end of English Mountain southwestward to the northeast end of Chilhowee Mountain, the northwest border of the Tellico formation is a thrust fault (Fig. 3) on which that formation rode over the Saltville block of the Great Valley. This fault is exposed on Little Pigeon River on the road that runs northwest from Richardson Cove. Southeast of the fault the road follows the gorge which Little Pigeon River and its tributary, Bird Creek, have cut through Shields Mountain (Fig. 3). The mountain is formed of thick, hard, closely folded quartzites and conglomerates that are the upper part of the Tellico formation. Where these quartzites on the west slope of Shields Mountain are exposed in the road cut, they are greatly crushed, rust-stained, and slickensided, have a steep southeast-dipping cleavage, and on their northwest side are in contact with sheared black shale, much mineralized by pyrite, and sandstone which comprise the Tellico sandstone of the Saltville block.

In the region north of Sand Mountain (Fig. 3) the boundary of the Tellico formation is very irregular and contorted and, 2 miles northeast of the fault contact on Little Pigeon River, passes around a westward-projecting hill called "Granny Knob" composed of quartzites and conglomerate similar to the rocks in Sand and Shields mountains. The rocks in "Granny Knob" are stratigraphically discordant to the formations adjacent to them. To the northeast, in a deep embayment in the trace of the fault, the overridden Knox dolomite extends up the flat valley of East Fork to the village of East Fork, and the overthrust beds of infolded quartzite and conglomerate of the upper part of the Tellico formation strike into, and are cut off by, the fault in the bluff on the south side of the embayment. Northeast of the embayment the fault curves around Little Mountain and passes under the Lower Cambrian quartzite of English Mountain.

In the Valley of Virginia the Pulaski overthrust is a fault of great magnitude that rides over the Saltville block. As shown on the map of the Appalachian Valley in Virginia (Butts, 1933), the overthrust trends southwest and passes into Tennessee west of Bristol. In Tennessee (Pond, 1933), the fault has been recognized to a point 6 miles southwest of Greeneville. The writers tentatively extend the Pulaski fault southwest from the vicinity of Greeneville to Newport, where a fault, marked by many springs, carries the Knox dolomite northwest over Athens shale (Fig. 3). They regard this fault and the fault on the north border of the Tellico formation from English Mountain southwestward to Chilhowee Mountain as the Pulaski thrust fault. The thrust passes under the quartzites at the north end of Chilhowee Mountain and emerges at the southwest end (Fig. 4). At Little Tennessee River impure limestone (Wilhite slate of Keith, 1896), in which the writers found fossils of Athens age, is thrust westward over sediments mapped as Bays sandstone in a syncline of the Saltville block.

The Pulaski overthrust emerges from beneath the Great Smoky block to the southwest at Rural Vale (Fig. 7), and the Tellico formation there is thrust over the Lower Cambrian Rome formation. The thrust passes under the overthrust Lower Cambrian quartzites of Starr and Beans mountains and is not exposed again to the southwest. At the southwest end of Beans Mountain the thrust evidently is hidden by the quartzites of Sugarloaf Mountain, for east of the mountain, at Parksville, a small area of the Tellico is exposed in a window. Farther southwestward the Pulaski thrust block underlies the Great Smoky and equivalent Cartersville block because the Tellico formation crops out in Alaculsy Valley, Ga., in a window in the Cartersville block (Pl. 1).

The writers' interpretation is that the Tellico formation was deposited southeast of its present position and, with underlying rocks, was thrust northwestward on the Pulaski overthrust. The Great Smoky overthrust overrode the Pulaski block so that only remnants of the Tellico formation are now exposed. Due to the increase in throw of the Great Smoky overthrust southwestward, the Great Smoky block covers the Pulaski block in southern Tennessee and Georgia, where the presence of the underlying Pulaski block is revealed only in the Alaculsy window. Inasmuch as the Pulaski and Saltville thrusts involve Carboniferous rocks, those thrusts took place during Appalachian orogeny, and the Great Smoky overthrust was somewhat later in the same orogeny.

AGE

The writers have found brachiopods in impure limestone (Wilhite slate of Keith, 1896) on Little Tennessee River just west of the south end of Chilhowee Mountain (Fig. 4). G. A. Cooper determined these fossils to be characteristic of the Athens shale of Middle Ordovician age. The senior author during earlier work southwest of English Mountain found in the vicinity of East Fork village (Fig. 3) graptolites in black fissile shale (Wilhite of Keith, Mt. Guyot folio) that resembles the Athens, and trilobites and brachiopods in buff sandy shale similar to the Tellico sandstone which overlies the Athens in the Saltville block northwest of English Mountain. The fossils were not specifically determined, but the graptolite-bearing shale is no doubt the same as that called Athens, and the fossiliferous sandy shale is Tellico sandstone. Although fossils have been found only in a few places in the shale, sandstone, and limestone here included in the Tellico formation, the writers regard all of them as of Ordovician age and as equivalent in part to the Tellico sandstone.

STRATIGRAPHIC EQUIVALENCE

The Tellico sandstone, as it has been mapped in the southeastern part of the Valley of east Tennessee, overlies the Athens shale and underlies the Sevier shale. Keith (1895b) named the Tellico sandstone from the section on Tellico River in Monroe County, Tenn., and described it as gray calcareous sandstone and sandy shale which weather to a red porous sandy rock. Farther south in Tennessee (Hayes, 1895) it is a bluish-gray sandy limestone which weathers to a red or purple porous sandstone. In most of the Valley of Virginia and the western part of the Valley of east Tennessee, including the western part of the Saltville block, the Middle Orodvician sediments are calcareous and finely argillaceous marine deposits in which the Tellico sandstone facies does not occur. This sandstone appears in southern Virginia in the Bristol (Campbell, 1899) and Abingdon quadrangles on the southeast side of the Pulaski block and is present also in that block in northeastern Tennessee east of Greeneville. The Tellico sandstone here overlies the Athens shale, is not fossiliferous, and is not overlain by fossiliferous formations; hence its exact upper limit is not known.

In the southeastern part of the Saltville block in Tennessee, where the Tellico sandstone was named, it is mostly a sandstone, and thick sandstone beds also occur in the overlying Sevier shale and in the underlying Athens shale and Knox dolomite. These sandstone beds were mapped (Keith, 1895b; 1896; Murphy folio; Hayes, 1895) as members of their respective formations and were said to represent a coarse clastic shore facies of deposition. In the syncline on the southeast side of the Saltville block (Fig. 4), the sequence below the Chattanooga shale (Keith, 1895b; 1896) is mapped as follows:

Bays sandstone
Sevier sandstone (lentil in Sevier shale)
Sevier shale
Tellico sandstone
Athens shale

G. A. Cooper (Personal communication, 1947) found fossils of high Ottosee equivalence at the base of the sandstone mapped as Bays. The Athens shale here also is

sandy. In this section the so-called Bays sandstone and underlying sandy beds in the Sevier, Tellico, and Athens represent shore facies of Middle Orodvician age deposited in the southeastern part of the basin while calcareous and finely argillaceous sediments were being deposited farther west. The exact stratigraphic position of the various sandy beds can be determined only by fossils. The lower part of the Tellico formation lithologically resembles the Tellico sandstone in the Pulaski and Saltville blocks in Virginia and Tennessee, and also resembles that sandstone in northern Georgia, where the Tellico sandstone as mapped contains thick beds of limestone conglomerate with pebbles of dove-colored limestone, quartz, and jasper.

The quartzose limestone, limestone conglomerate, and interbedded calcareous shale in the lower part of the Tellico formation therefore may be equivalent in time to part of the Athens shale, Chickamauga limestone, and possibly the upper part of the Knox dolomite. Where the Tellico formation overlies the Athens shale, it represents the Tellico sandstone and higher beds. The arkosic conglomerate and white quartzite that overlie the beds believed to be equivalent to the Tellico sandstone are younger beds in the clastic shore facies whose age is not definitely known. The writers conclude, therefore, that the Tellico formation, composed of limestone, shale, and sandstone, formerly mapped with Wilhite and Hiwassee slates, is the southeastward expansion of clastic beds which were deposited in the southeastern part of the Pulaski and Saltville blocks, and may represent (Fig. 11) the shore facies of all the Ordovician formations above the Knox dolomite; that this shore facies was deposited more or less synchronously with the marine, calcareous, and finer argillaceous sediments which accumulated in the deeper off-shore parts of the basin in the western part of the Saltville block.

LOWER CAMBRIAN QUARTZOSE ROCKS

GENERAL STATEMENT

The Lower Cambrian quartzose rocks that underlie Lower Cambrian dolomite occur on the flanks of uplifts of pre-Cambrian rocks and extend from Pennsylvania to Alabama. Some geologists (Resser, 1938, p. 2) have maintained that the base of the Lower Cambrian should be placed directly beneath the fossiliferous beds in the uppermost quartzite (Hesse, Erwin, or Antietam), and that underlying quartzites and shale should be placed in the pre-Cambrian. However, the quartzose rocks are conformable from the fossiliferous beds down to the pebbly arkose that initiates the series, and logically all of these conformable rocks should be part of the Lower Cambrian.

The quartzose rocks of this conformable Lower Cambrian sequence in places unconformably overlie early pre-Cambrian granite gneisses and highly metamorphosed sediments which make up an injection complex. At other places these quartzites overlie effusive and clastic rocks—the Catoctin basalt, Swift Run tuff, Lynchburg gneiss, Mt. Rogers volcanic series, and Ocoee series. They in turn unconformably overlie the injection complex. Some geologists (Bloomer and Bloomer, 1947) have maintained that these effusive and clastic rocks beneath the Lower Cambrian quartzose rocks are Lower Cambrian. The writers present evidence that the conformable sequence of Lower Cambrian rocks transgresses the underlying effusive and clastic

rocks and unconformably laps over them, and therefore these effusive and clastic rocks are late pre-Cambrian.

LITHOLOGY

The Lower Cambiran quartzose rocks from Pennsylvania southwestward have a marked lithologic similarity. They vary somewhat in lithologic aspect and stratigraphy across the strike because they were deposited at different distances from the shore, and along the strike because they were deposited in different parts of the sea under varying environments and conditions. The Lower Cambrian quartzose rocks on the west flank of the Blue Ridge and the Blue Ridge Plateau have three distinct facies. Where these facies are adjacent, they were brought into juxtaposition by thrust faulting (Fig. 6). The facies recognized are (1) the Chilhowee group in Tennessee, and similar quartzites in the Pulaski block in Virginia; (2) the facies in the Holston Mountain block; and (3) the facies in the Blue Ridge-Catoctin Mountain anticlinorium.

Formations of the Chilhowee group and of the Holston Mountain block are compared in Table 4, which is a revision of the correlation given by the writers in their previous paper (Stose and Stose, 1944, p. 389).

The section of the Chilhowee group in Chilhowee Mountain (Fig. 4) exposed on the State Highway along Little River (Knoxville quadrangle) is as follows:

Section of Chilhowee group (facies I) west of Walland, Tennessee

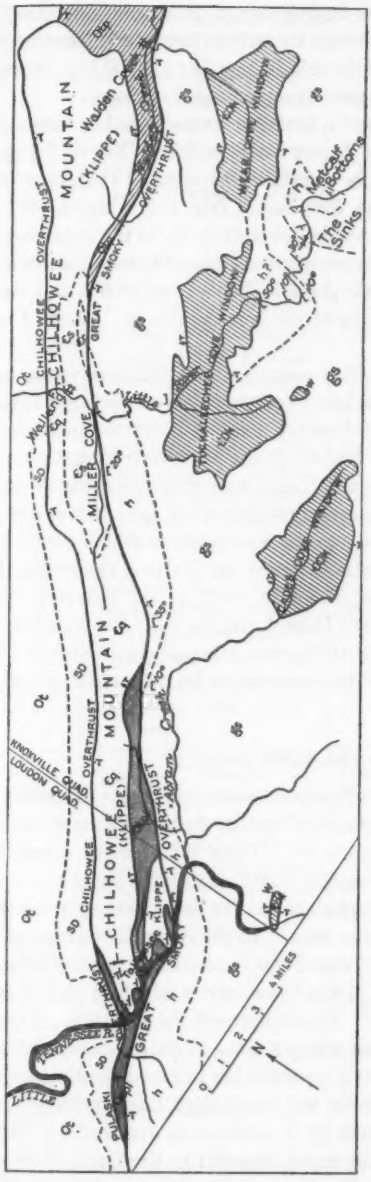
		Feet
Shady dolomite (Knox dolomite of Keith, 1895 b)		
Thick-bedded knotty dolomite		
Hesse quartzite		
White and banded quartzite containing round grains of glauconite and interbedded shale	20±	
Thin-bedded hard fine-grained gray quartzite with interbedded black shale	110	
Thick-bedded gray to white hard vitreous quartzite, *Scolithus*-bearing	120	250
Murray shale		
Buff shale, poorly exposed		500±
Nebo quartzite		
Thick-bedded white quartzite		
Cochran quartzite		
Thick-bedded hard coarse arkosic quartzite with conglomerate beds composed of rounded quartz pebbles		330
Thin-bedded green quartzite		
(Lower beds not exposed due to faulting)		

The writers have recently established the normal stratigraphic relations of these Lower Cambrian quartzose beds to the dolomite and red shale that underlie the Miller Cove around Walland (Fig. 4). The uppermost quartzite exposed at Walland is calcareous, contains glauconite grains and shale partings, and represents the characteristic transition zone at the top of the equivalent Erwin quartzite in Shady Valley, in Stony Creek valley (Stose, 1942, p. 167–172), and elsewhere. The Lower Cambrian fossils collected at Walland by Walcott (Keith, 1895b) probably came from these transition beds which are commonly fossiliferous. Red shale, which crops out northeast of Walland and was mapped as shore deposits of the Knox, is Rome formation

TABLE 4.—*Comparison of the Chilhowee group (facies 1) and the Lower Cambrian quartzose formations of the Holston Mountain block (facies 2)*

	a		b	c
	Facies 2, in Holston Mountain block, Roan Mountain quadrangle northeastward		Facies 1, Chilhowee group in Hot Springs window	Facies 1, Chilhowee group in Cherokee, Meadow Creek, Chilhowee, Beans, and Starr mountains (Keith, 1895b; 1896; 1904; 1905a; Hayes, 1895)
Lower Cambrian	Erwin quartzite		Hesse quartzite	Hesse quartzite
	Hampton shale		Murray shale	Murray shale
	Unicoi formation	Upper member (hard white quartzite; current-bedded green to dark banded in lower part)	Nebo quartzite (hard white current-bedded quartzite)	Nebo quartzite (hard white quartzite)
				Nichols shale (thin shale and quartzite)
		Middle member (basalt flows and interbedded shale and arkose)	Cochran quartzite (coarse arkosic quartzite and conglomerate; includes Nichols shale of Keith)	Cochran conglomerate (arkosic quartzite and conglomerate)
		Lower member (coarse arkose and conglomerate; quartz pebble bed at base)		
			Sandsuck shale (shale and soft arkose)	Sandsuck shale (thick shale and arkose; Starr conglomerate locally) (base of Cambrian not exposed)
Late pre-Cambrian	Mt. Rogers volcanic series (in places)		Vann quartzite (arkosic quartzite and soft arkose)	
Early pre-Cambrian	Injection complex		Injection complex	

overlying the dolomite which is Shady. In the Chilhowee group the quartzite called Nebo, just below the Murray shale, is a thick, massive, white, current-bedded quartzite, similar to the quartzite in the upper part of the Unicoi formation of the Holston Mountain sequence. The Cochran quartzite, just below the Nebo, is an arkosic quartzite similar to part of the lower member of the Unicoi. The lowest beds exposed in the Chilhowee group are soft fine-grained arkose and shale (Sandsuck formation, which has no known counterpart in the Unicoi). The term Unicoi formation (Table 4) is applied to all the quartzose rocks of facies 2 below the Hampton shale.



Great Valley sequence
SD, Silurian and Devonian
Ot, Tellico sandstone (after Keith)
€Ok, Knox dolomite (in windows)
Pulaski overthrust block
Otp, Tellico formation
Otpl, Limestone and shale at base of Tellico formation
Chilhowee Mountain thrust block (klippe)
€r, Rome formation
€s, Shady dolomite
€q, Lower Cambrian quartzose rocks
Great Smoky overthrust block
gs, Great Smoky quartzite } Ocoee series
h, Hurricane graywacke }

F, Fossils
T, Overthrust side of thrust fault
T↑, Direction of dip of thrust fault on east side of klippe and west side of window

FIGURE 4.—*Geological map of vicinity of Chilhowee Mountain*

Showing Tellico formation overridden by Ocoee series in Great Smoky block and by Lower Cambrian quartzites in the Chilhowee Mountain klippe; windows in the Great Smoky block expose Knox dolomite. Adjoins Figure 3 on south.

The Unicoi is divided into three members, the middle one of which contains basalt flows. In a well-exposed section of the Unicoi which the writers measured in southern Virginia (Jonas and Stose, 1943b) the flows of the middle member are 700 feet above the base of the Lower Cambrian. The Chilhowee group contains no flows.

Lower Cambrian quartzites that do not contain basalt flows and in other respects resemble the Chilhowee group crop out in the Taylor Valley window (Fig. 6) (Stose and Jonas, 1938, p. 23). The section of these quartzites exposed in the gorge of Doe River southeast of Hampton, Tenn. (King, Ferguson, Craig, and Rodgers, 1944, Fig. 6, Sec. I), is complete from the overlying Shady dolomite down to the basal beds which rest on the injection complex. Another complete section of Lower Cambrian quartzites that resemble the Chilhowee group is given in Table 5, column a, and the variation of this facies northeastward along the strike is in column b. Facies 3 is shown in column c.

The formation names applied to quartzites that resemble the Chilhowee group in the Buena Vista anticline and Fullhardt Knob block (column a) are those that have been previously used and are tentative. The formation called Weverton quartzite does not contain dark ferruginous quartzite typical of the Weverton formation in facies 3 (column c). The basal beds of quartzite (column a) resemble those of the Loudoun formation of facies 3 in that they are made up of thin flows, tuffs, and arkose, and differ from the Chilhowee group (facies 1) to the southwest in which volcanic rocks are not present at the base. The quartzites below the Harpers shale should not be called Unicoi formation (Bloomer and Bloomer, 1947, p. 96, 103–106) for they do not contain flows in the middle member (Table 4, column a). The *Scolithus*-bearing quartzite member in the Harpers shale thickens in Pennsylvania (column b) where this member (Montalto) largely takes the place of the Harpers shale (Stose, 1910, p. 30–38).

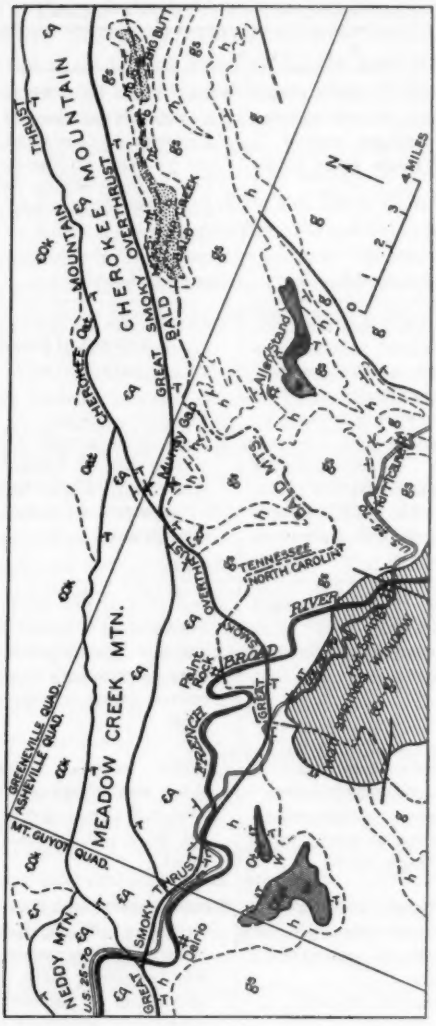
STRUCTURAL RELATION TO THE OCOEE SERIES

The Chilhowee group, overlain in places by Shady dolomite and Rome formation, is in a series of thrust slices in Cherokee and Meadow Creek mountains, in the vicinity of Paint Rock, and in Stone Mountain (Figs. 3, 4, 5). These thrust slices override younger rocks of the Pulaski block. The bounding overthrusts are a series of overlapping en echelon thrusts, and each slice is terminated at the southwest by a thrust fault which overrides the succeeding slice to the west. In the Asheville quadrangle and vicinity, Keith (1904) did not map the Great Smoky overthrust, or any other major thrust fault, on the northwest border of the Ocoee series which in that area he regarded as part of the Chilhowee group. He interpreted the structure of the northern part of the Ocoee series as thrust slices closely related to the imbricate thrust slices of the Chilhowee group. However, the Ocoee series lies in a long synclinorium along the northwest border of the Great Smoky and equivalent Cartersville overthrust block and is bounded on the northwest by a continuous overthrust. The bordering faults of the slices mentioned above trend diagonal to the Great Smoky overthrust, and that fault cuts off the slices on their southeast sides (Figs. 3, 5). Where the Great Smoky overthrust cuts off the southwest end of the western slice (Stone Mountain) the fault is exposed in a quarry ½ mile south of Cosby (Fig. 3). In the quarry, closely folded phyllite of the Hurricane graywacke overlies the Hesse

TABLE 5.—*Lower Cambrian quartzose facies in Virginia and northward into Pennsylvania*

	a	b	c
	Facies 1, in Buena Vista anticline, Va., west of Montebello thrust, and Fullhardt Knob block north of Roanoke (Fig. 6)	Facies 1, in South Mountain, Pa., and northern Maryland, west of Harpers Ferry thrust	Facies 3, in Blue Ridge-Catoctin Mountain anticlinorium, Md. and northern Va., east of Harpers Ferry thrust; central Va. east of Montebello thrust (Fig. 6)
Lower Cambrian	Tomstown dolomite	Tomstown dolomite	Tomstown dolomite
	Antietam (Erwin) quartzite (thick-bedded *Scolithus*-bearing quartzite, fossiliferous at top).	Antietam quartzite (thick-bedded *Scolithus*-bearing quartzite, fossiliferous at top).	Antietam quartzite (present only at east foot of Catoctin Mtn., Md.)
	Harpers (Hampton) shale, containing a *Scolithus*-bearing medial member (Snowden member of Bloomer).	Harpers shale, containing a *Scolithus*-bearing medial member (Montalto), which replaces most of the shale at the north	Harpers shale (present only at east foot of Catoctin Mtn., Md.). Higher beds not present in synclines of anticlinorium.
	"Weverton" quartzite (Unicoi formation of Bloomer) (massive white and thin-bedded green quartzites, argillite, and pebbly arkose)	Weverton quartzite (coarse hard gray quartzite and conglomerate)	Weverton quartzite (white, ridge-making quartzite, arkose, purple-banded quartzite with dark glassy quartz grains in upper part)
	"Loudoun" formation in places (arkosic quartzite, red to purple spotted tuff, thin andesite flows, and coarse arkose at base)	Loudoun formation (soft shale, pebbly arkose, and conglomerate; some tuffaceous beds at base in places)	Loudoun formation (thin shale and conglomerate beds; tuff, thin amygdaloidal flows, and arkose at base)
Late pre-Cambrian		Rhyolite and Catoctin basalt flows, rhyolite tuff and agglomerate at top. (Base not exposed)	Catoctin basalt (thick flows). Swift Run tuff (tuff, shale, flows, and arkose; equivalent Lynchburg gneiss on east flank of anticlinorium)
Early pre-Cambrian	Injection complex		Injection complex

quartzite in overthrust relation, and the thrust contact dips 35° W. and strikes nearly due north. This divergence from the east-west trend of the fault in that vicinity suggests that the thrust plane there is locally deflected.



Great Valley sequence
Oat, Athens shale and Tellico sandstone
€Ok, Knox dolomite
€r, Rome formation
Meadow Creek and Cherokee mountains fault slices
€q, Largely Lower Cambrian quartzose rocks; some Shady dolomite and Rome formation
€r-g, Rome formation to pre-Cambrian granite in Hot Springs window.
Pulaski thrust block in windows
Otpl, Limestone and shale at base of Tellico formation
Great Smoky overthrust block
b, Big Butt quartzite
nt, Nantahala slate
gs, Great Smoky quartzite
h, Hurricane graywacke
} Ocoee series
g, Granite and gneiss, injection complex

T, overthrust side of thrust fault
T↑, Direction of dip of thrust fault

FIGURE 5.—*Geological map of vicinity of Cherokee and Meadow Creek mountains*
Showing Lower Camb.ian quartzite slices in front of Great Smoky thrust and windows in Great Smoky thrust block.

South of the Paint Rock slice the quartzites of the Chilhowee group, exposed in the Hot Springs window (Stose and Stose, 1947) surrounded by the Ocoee series in the Great Smoky block, are part of a slice structurally related to those that lie northwest of the Great Smoky overthrust.

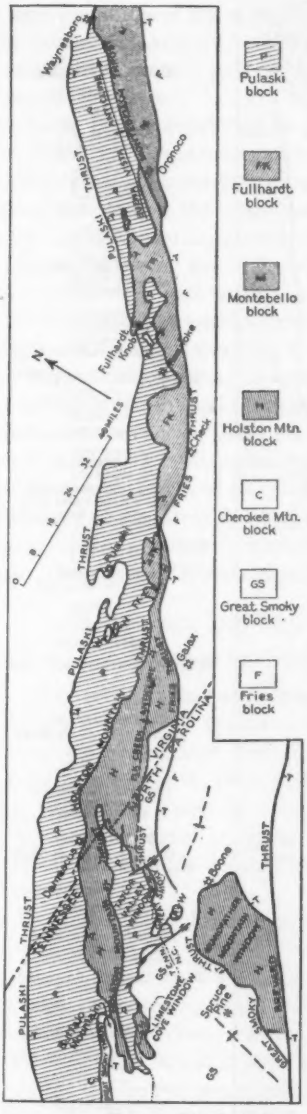

Waynesboro
Pulaski block
Fullhardt block
Montebello block
Holston Mtn. block
Cherokee Mtn. block
Great Smoky block
Fries block
Injection complex overlain by Lower Cambrian facies similar to Chilhowee group; in Buena Vista anticline.
Injection complex overlain by Lower Cambrian facies same as that in Buena Vista anticline.
Injection complex overlain by late pre-Cambrian Swift Run tuff, Catoctin basalt, and Lower Cambrian facies with Loudoun formation and Weverton quartzite; volcanic beds at base of Loudoun.
Injection complex overlain by late pre-Cambrian Mt. Rogers volcanic series, Lower Cambrian facies with Unicoi formation; overlaps volcanic series and in places lies on injection complex.
Lower Cambrian, Chilhowee group facies.
Injection complex overlain by Ocoee series; Unicoi formation overlies Ocoee series in gorge of Nolichucky River.
Injection complex overlain by Lynchburg gneiss
K, Klippe
W, Window
T, Overthrust sides of thrust fault
T↑, Direction of dip of thrust faults

FIGURE 6.—*Tectonic map of part of the Appalachian Mountains*
Showing fault blocks and character of formations in each block.

The Chilhowee group in English, Chilhowee, Starr, and Beans mountains (Figs. 3, 4, 7) are klippen that in places cover the trace of the Pulaski overthrust and are

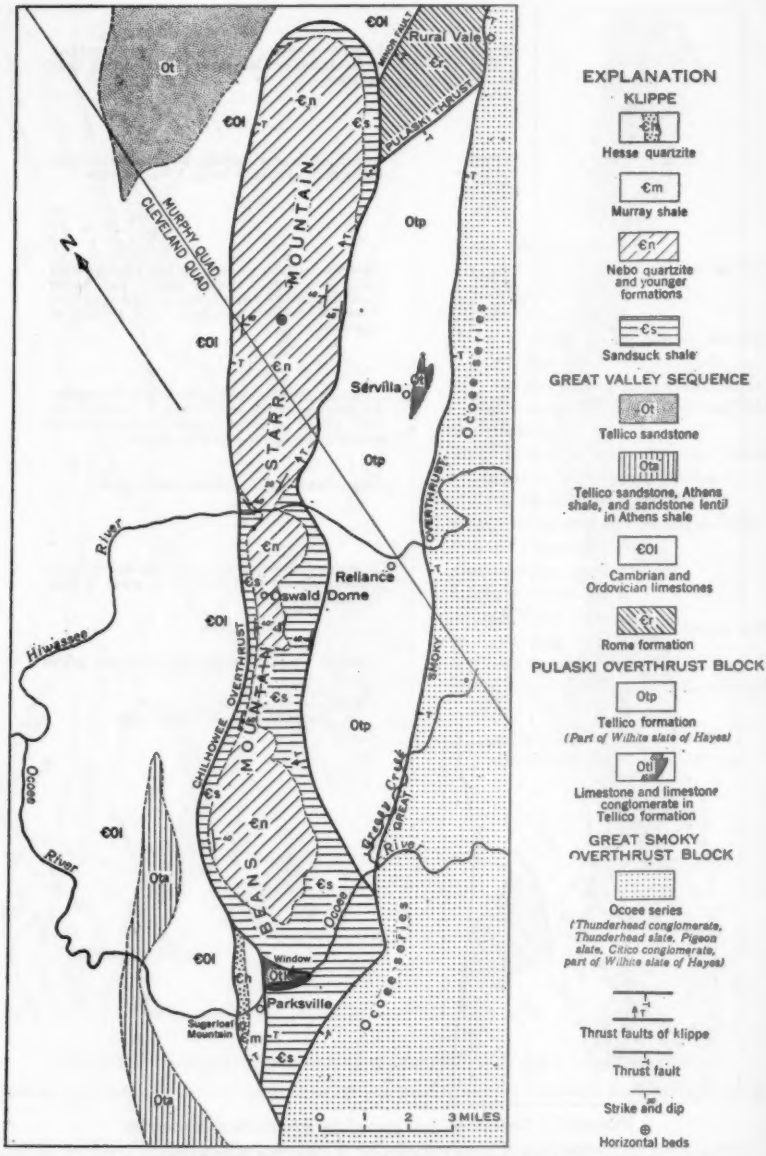


FIGURE 7.—*Generalized geological map of Starr and Beans mountains klippe, southern Tennessee.*

thrust over the edge of the Saltville block. On their southeast sides they overlie in thrust relation the Tellico formation. At the southwest end of Chilhowee Mountain, north of Little Tennessee River (Fig. 4), where this relation is well shown, the Lower

Cambrian quartzites of the mountain are cut through by tributary streams, and the Tellico formation is exposed in the valleys. Near Tallassee, limestone of the Tellico formation at the contact with the quartzites is brecciated and infiltrated with vein quartz and iron oxide.

The Great Smoky overthrust in places overrides the southeast side of the Chilhowee, Starr, and Beans mountains klippen, and in southern Tennessee cuts off the southwest end of the latter klippe and rides over the Saltville block (Fig. 7). Overthrusting has brought the Ocoee series in the Great Smoky block in juxtaposition with the Lower Cambrian quartzites of the Chilhowee group. As the thrust slices and klippen containing the Chilhowee group are thrust over the Pulaski block, the faults on which they rode northwestward are not the Pulaski overthrust. Because the Chilhowee group resembles Cambrian quartzites in the Pulaski block exposed in the Taylor Valley window and in the Buena Vista anticline (Table 5, column a), the writers suggest that the quartzites of the klippen and thrust slices originated in anticlinal uplifts in the Pulaski block and were sheared off and pushed in front of the northwestward-advancing Great Smoky overthrust block and rode over higher beds of the Pulaski block. According to this interpretation of the structure, the Chilhowee group and the Ocoee series were not deposited in the same thrust block, and the thrusting of the Great Smoky block and of the slices containing the Chilhowee group occurred in the Appalachian orogeny, later than the westward advance of the Saltville, Pulaski, and Holston Mountain thrust blocks.

AGE AND RELATIONS OF OCOEE SERIES

COMPARISON OF OCOEE SERIES WITH CHILHOWEE GROUP

The Ocoee series at first was regarded as entirely distinct in lithology from other sedimentary rocks in the Appalachians, particularly different from the Lower Cambrian quartzose formations (Safford, 1856; Keith, 1895b; 1896; Hayes, 1895). In some later reports (Keith, 1905a; 1907a; 1907b; 1907c; La Forge and Phalen, 1913; Bayley, 1928; King, Ferguson, Craig, Rodgers, 1944, Pls. 1, 8) rocks of the Ocoee series have been mapped as Lower Cambrian. The writers show that the Ocoee series does not resemble lithologically the Chilhowee group or any other Lower Cambrian facies.

The Chilhowee group, in common with other facies of the Lower Cambrian quartzites, is well bedded and consists of well-sorted material—smoothly rounded pebbles and grains of quartz with a subordinate amount of feldspar grains in a matrix of quartz sand, cemented by silica. The lower part of these quartzites contains conglomerate made up of clean, rounded, quartz pebbles in a matrix of firmly cemented sand (Pl. 4, fig. 1). Keith (1895b; 1896) does not mention the occurrence of graywacke in the Chilhowee group, and none is present in that group. Adolph Knopf determined the specific gravity of a specimen of compact white quartzite from the west crest of Beans Mountain, Tenn., to be 2.64, which is that of quartz. These pure white quartzites lie both above and below the horizon of the Hampton shale; they show current bedding; the upper beds contain *Scolithus linearis* and have at the top Lower Cambrian fossils. The Lower Cambrian quartzites persist with regular thickness along the strike and form long straight strike ridges.

In contrast the Ocoee series are much more variable in character than the Lower Cambrian quartzites, as expressed by Hayes (1895, p. 2): "The formations (of the Ocoee series) have so few constant characteristics and undergo so many variations, even with the most careful work one may go far astray in making correlations." The Ocoee series is made up of dominantly dark, dense, fine-grained to coarse-grained graywackes. The grains and pebbles are angular and comprise, besides quartz and feldspar, granodiorite and granite gneiss, and black slaty graywacke. The dark color of the graywacke and slate is due to the presence of iron oxide, carbonaceous material, and metamorphic biotite, and does not disappear on weathering. The quartzites in the Great Smoky and the Big Butt quartzites are largely devoid of bedding planes and occur in layers up to 20 feet thick without visible parting planes. Except in the siliceous-banded beds the Hurricane also shows no well-defined layering. Hard beds do not persist for great distances along the strike, and their topographic expression is in short ridges and domelike peaks. The formations vary in thickness; the variation is most pronounced in the Great Smoky quartzite—estimated minimum 500–800 feet in the northeastern part of the area to a maximum of 4100 feet in the Great Smoky Mountains in the vicinity of Newfound Gap. The Big Butt quartzite is thickest in the northeast part of the area, is thinner in the northeastern part of the Murphy syncline, and disappears southwestward along the strike in Georgia. The pebbles of fine-grained, black, slaty graywacke in the middle and upper members of the Great Smoky quartzite resemble graywacke of the Hurricane and probably were derived from older beds of the Ocoee series which were locally uplifted and eroded during its deposition. The angularity and mineralogy of most of the detrital material in the series show that the fragments were derived from the rapid disintegration of granitic rocks similar to the early pre-Cambrian injection complex in near-by elevated areas. The poor stratification and variable thickness of layers indicate that waste from the bordering highlands was supplied irregularly by streams that did not have steady and continuous currents, and that the sediments were deposited in a continental basin, perhaps of local extent, formed by downwarping of the early pre-Cambrian injection complex on which the series lies unconformably. The abundant carbonaceous material in the Ocoee series probably represents primary carbon in the sediments and may have formed as bituminous mud in a deep, closed basin. It may be positively stated therefore that the Chilhowee group and other facies of Lower Cambrian quartzites are not similar to the Ocoee series in lithologic character or in manner of deposition. Since they were deposited in different thrust blocks, it might be claimed that their dissimilarity is not necessarily proof of a difference in age.

METAMORPHISM AS EVIDENCE OF RELATIVE AGE

The Lower Cambrian rocks of the Chilhowee group that lie northwest of the Ocoee series are not metamorphosed. The rocks of the Ocoee series are greatly indurated, contain a variable amount of metamorphic muscovite and biotite, and in the southeastern part of the belt of the series, from the southern part of the Mt. Guyot quadrangle southwestward, are highly metamorphosed and contain staurolite, garnet, and kyanite. Especially in the vicinity of thrust faults that bound the Murphy syncline,

the earlier-formed structures are deformed, bedding in the softer beds is in large part obliterated, and minerals of high-rank metamorphism have been reduced in metamorphic rank showing that the series has experienced two periods of deformation and metamorphism. Two periods of metamorphism have been mentioned also by previous writers (La Forge and Phalen, 1913, p. 10; Bayley, 1928, p. 54, 66; Crickmay, 1936, p. 1389–1390). All assigned the later deformation, the metamorphism, and the thrust faulting in the Murphy syncline to post-Carboniferous orogeny. La Forge and Phalen, who believed that the Ocoee series is Lower Cambrian thought (1913, p. 10) that the earlier folding and metamorphism occurred at the close of the Ordovician period. Crickmay (1936, p. 1392) concluded that the Talladega slate in Georgia (Ocoee series) was first folded and metamorphosed in the pre-Cambrian because the adjacent Lower Cambrian rocks are not metamorphosed.

In the Elk Creek anticline (Fig. 6) the Mt. Rogers volcanic series, regarded by the writers as of late pre-Cambrian age, is only gently folded. It has been hydrothermally altered but is only slightly affected by dynamometamorphism as is shown by the fact that the volcanic series contains soft siliceous-banded shale and water-laid tuff in which ripple marks (Pl. 4, fig. 2) are preserved. In northern Virginia and southern Maryland the Catoctin basalt and Swift Run tuff, also regarded by the writers as of late pre-Cambrian age, are closely folded and have a strongly developed cleavage, but the structures conform to those in the overlying Paleozoic rocks and were formed simultaneously in both rock series during late Paleozoic orogeny.

The Murphy marble in the Murphy syncline, in common with the underlying Ocoee series in the vicinity of the syncline, is metamorphosed and contains such minerals as garnet, tremolite, phlogopite, and biotite. Along fault zones tremolite has been sheared to talc. If the writers' assumption is correct that the Murphy marble is of Lower Cambrian age, the metamorphism of the Murphy marble and the neighboring Ocoee series is post-Cambrian. West of the Grandfather Mountain window, in the vicinity of Spruce Pine, N. C. (Fig. 6), the Lynchburg gneiss and intercalated basic sills of late pre-Cambrian age are intruded and injected by aplitic granite and mica-bearing pegmatite (Olson, 1944, p. 22–36) which are little deformed, whereas the rocks they intrude are closely folded and metamorphosed. The mica-bearing pegmatite entered these rocks therefore at the close of the period of folding. A crystal of uraninite from pegmatite near Spruce Pine was analyzed (Alter and McColley, 1942, p. 213) in three zones. The age calculated from the determination of the lead-uranium-thorium ratio in the different zones ranges from 358 to 382 million years, which would place the age of the pegmatite as Ordovician. These results are in close agreement with the analysis by Hillebrand (1891, p. 65–67) of uraninite in pegmatite from the Flat Rock mine near Spruce Pine, the calculated age of which, when corrected for actinium D, (Adolph Knopf, personal communication, 1947) is 350 million years. Because the isotopes of lead in the uraninite were not determined, Knopf says it is not possible to calculate the ratio with certainty, so that the Ordovician age of the pegmatite near Spruce Pine is not fully established.

Mica-bearing pegmatite similar to that near Spruce Pine intrudes the southeast side of the belt of the Ocoee series from the south side of the Mt. Guyot quadrangle southwestward across southern North Carolina and northern Georgia, where the

metamorphism of the Ocoee series is of high rank; there is no pegmatite where the metamorphism of the series is of low rank. The coextent of high-rank metamorphism and of the pegmatitic intrusions suggest that the localization of high-rank metamorphism in the Ocoee series may have been largely due to thermal solutions that accompanied the pegmatite. If the mica pegmatite in the Ocoee series is equivalent in age to the pegmatite near Spruce Pine, the Ocoee series was apparently folded in Ordovician orogeny and the main synclinal structure of the series was then formed; the series as a whole was mildly metamorphosed, and the part of it in the vicinity of the pegmatite intrusions was highly metamorphosed.

In Appalachian orogeny the Ocoee series, with low-rank metamorphism along the Great Smoky overthrust, was crumpled and sheared into phyllite without change in metamorphic rank. Along the subsidiary thrusts in the Great Smoky block, such as the Whitestone and Tusquitee thrusts (Pl. 1) which occurred during or just after the main overthrusting, the highly metamorphosed parts of the Ocoee series were made schistose, bedding was in large part obliterated, and high-rank minerals were degraded in metamorphic rank.

If the pegmatite in the Ocoee series is Ordovician, the first metamorphism and folding of the series occurred during Taconic orogeny and long preceded the second deformation. If the pegmatite is late Paleozoic, all the metamorphism and folding of the Ocoee series took place in stages of Appalachian orogeny. The reasons cited above rule out the possibility that the folding and metamorphism of the Ocoee series are pre-Cambrian, and the fact that the Ocoee series is more metamorphosed than the Paleozoic rocks which lie northwest of it is not a valid argument for a greater age for the Ocoee series as concluded by Crickmay (1936, p. 1392).

ROCKS OF LATE PRE-CAMBRIAN AGE

The core of the Blue Ridge-Catoctin Mountain anticlinorium (Jonas and Stose, 1939, p. 576–579, Fig. 1) is made up of metamorphosed igneous and sedimentary rocks which form an injection complex of early pre-Cambrian age. At their northeast-plunging end in Maryland they are exposed in the Middletown and Rohrersville anticlines which merge southward in Virginia. The injection complex is overlain, in rising sequence, by the Lynchburg gneiss or the Swift Run tuff (Stose and Stose, 1946a, p. 18–19), Catoctin basalt, and the Lower Cambrian Loudoun formation and Weverton quartzite (Table 5, column c). The Lynchburg gneiss on the east side of the anticlinorium, at its type locality, Lynchburg, Va., and southwestward, is a mica gneiss and mica schist. In northeastern Virginia and Maryland, where it is less metamorphosed, the formation is graphitic slate, sericite schist, and sericitic arkosic quartzite. It thins in that area and contains the Goose Creek marble (Stose and Stose, 1946a, p. 19–20) which is interbedded with basalt flows. The Lynchburg gneiss passes northward along the strike into the Swift Run tuff, composed of tuffaceous slate, thin flows, and arkose, which similarly overlie the injection complex and underlie the Catoctin basalt. The Lynchburg gneiss, therefore, is equivalent to the Swift Run tuff. The Lynchburg gneiss rests on the complex with a coarse granitic conglomerate (Rockfish) at its base (Nelson, 1932; Stose and Stose, 1944, p. 409) and is clearly younger than the injection complex. The gneiss is intruded by basic

sills related to the Catoctin basalt. As these sills do not penetrate the Lower Cambrian quartzose rocks that overlie the Catoctin basalt, the basalt and the Lynchburg gneiss are older than Lower Cambrian. The Catoctin basalt and the underlying Swift Run tuff pass around the north end of the Middletown anticline and extend southward into Virginia where they are almost continuously exposed along the east side of the Blue Ridge to the south end of the Montebello block at a point south of Oronoco, where that block is overridden by the Fries overthrust (Fig. 6). Throughout that area these coextensive formations overlie the injection complex and underlie the Lower Cambrian Loudoun formation and Weverton quartzite.

In central Virginia a series of flows, tuffs, and arkoses has been described as the Oronoco formation (Bloomer and Bloomer, 1947, p. 94–97, 102–106). At the type locality this formation overlies the injection complex, is overlain by the Catoctin basalt and is equivalent to the Swift Run tuff which had previously been named and mapped as far south as this part of Virginia (Stose and Stose, 1946, p. 18–19; Jonas and Stose, 1939, Fig. 1), so that a new name for the formation is not warranted.

In their survey of part of Clarke County and adjoining Loudoun County, northern Virginia, the writers (Jonas and Stose, 1943 a) found evidence of broad pre-Cambrian doming and erosion of the Catoctin basalt (Fig. 8). In that area the Catoctin forms the crest and west slope of the Blue Ridge and lies southeast of a band of Loudoun formation and overlying Weverton quartzite which crop out in foothills and linear ridges at the northwest border of the mountain slope. There the Loudoun formation, at the base of the Lower Cambrian, directly overlies the Catoctin basalt, and the Catoctin basalt, with the Swift Run tuff at its base, overlies the injection complex of early pre-Cambrian age. At Bluemont, east of Snickers Gap, the Swift Run tuff and the overlying Catoctin basalt bend sharply westward in a domal uplift that exposes the underlying injection complex 1 mile west of the general trend of its outcrop. West of this anticline the Lower Cambrian Loudoun formation and Weverton quartzite trend in a straight line and are not affected by the doming of underlying rocks at Bluemont, and the Catoctin basalt here is thinner. At Purcell Knob, 8 miles northeast, and northeastward to Potomac River, the basalt and the Swift Run tuff are absent, and the Loudoun formation overlaps them, resting directly on granodiorite of the injection complex. In this region Keith (1895a, p. 324–329) mapped as Loudoun formation lenticular areas of arkose at the contact of the Catoctin basalt and granite gneiss (injection complex). These arkoses however are part of the Swift Run tuff. Thin flows, tuffs, and arkose also occur at the base of the Loudoun formation and were recognized and mapped by the writers in Maryland (Stose and Stose, 1946a). Near Purcell Knob, where the Loudoun formation overlaps the Swift Run tuff, the basal beds of the Loudoun comprise thin flows, tuffs, and arkose. Similar beds have been described (Furcron and Woodward, 1936, p. 47–51) farther south in Virginia. The writers ascribe the thinning of the Catoctin basalt at Bluemont and its absence at Purcell Knob and northeastward to erosion that followed an uplift which took place in the late pre-Cambrian before the Lower Cambrian was deposited.

In Virginia, on the north side of the Elk Creek anticline (Jonas and Stose, 1943b),

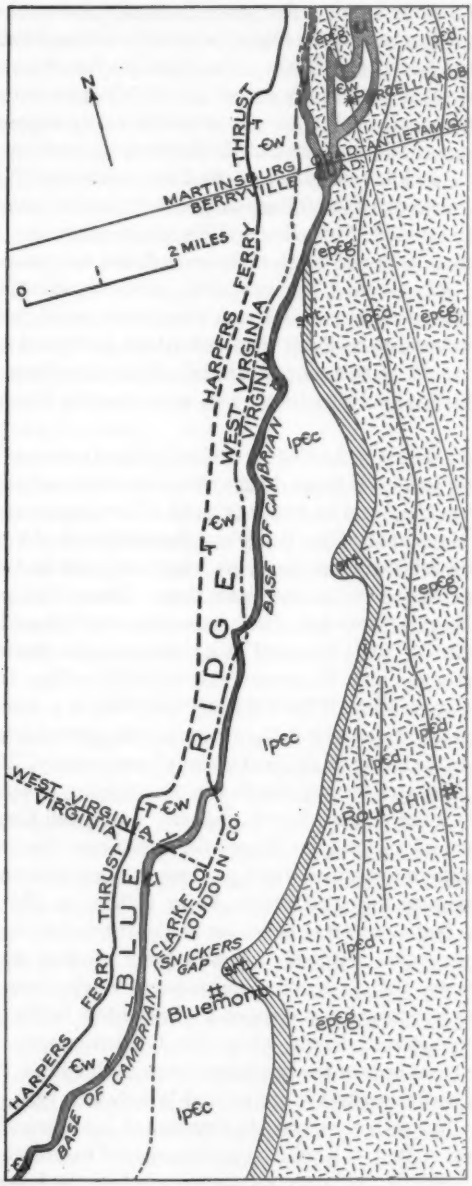


FIGURE 8.—*Geological map of the Blue Ridge and vicinity in northern Virginia*

Showing unconformable relations of the base of the Cambrian to underlying formations. €w, Weverton quartzite. €l, Loudoun formation with volcanic beds at base. lp€c, Catoctin basalt. lp€d, diabase dikes related to Catoctin basalt. srt, Swift Run tuff. ep€g, granitic rocks of early pre-Cambrian injection complex.

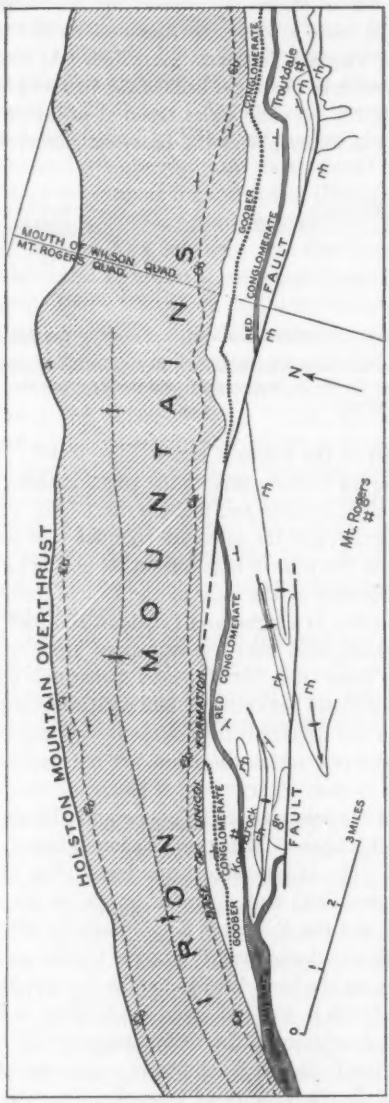


FIGURE 9.—*Geological map of area north of Mt. Rogers, Virginia*

Showing unconformable relations of the base of the Cambrian (Unicoi formation) to underlying Mt. Rogers volcanic series. Ruled pattern, Lower Cambrian quartzites; Cb, basalt member of Unicoi formation. "Goober" quartz conglomerate and red conglomerate containing granite boulders in Mt. Rogers volcanic series were truncated by erosion prior to deposition of the Cambrian. rh, rhyolite flows in Mt. Rogers volcanic series. gr, granite of injection complex.

the Lower Cambrian Unicoi formation overlies the Mt. Rogers volcanic series for 30 miles (Figs. 6, 9, 10), from a point near the west border of the Speedwell quadrangle southwest to the Virginia-Tennessee State line. At no outcrop of this contact is there a marked difference in dip of the beds of the two series, but along the contact as a whole the beds of the volcanic series trend at an angle with the strike of the Unicoi formation, and in many places the basal conglomeratic beds of the Unicoi

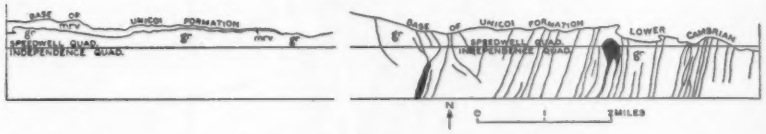


FIGURE 10.—*Geological map of southwest and southeast edges of the Speedwell quadrangle, Virginia*

Showing at the left, the base of the Cambrian overlapping the Mt. Rogers volcanic series (mrv), and at the right, swarms of rhyolite and basalt dikes related to the Mt. Rogers volcanic series; dikes intrude the granite of the injection complex (gr) and do not penetrate the Cambrian.

overlap mappable units of the volcanic series (Figs. 9, 10). In the eastern part of the Elk Creek anticline the Unicoi overlaps the volcanic series and rests on the injection complex. Swarms of rhyolite and diabase dikes (Fig. 10), syngenetic with the Mt. Rogers volcanic series, cut the injection complex just south of the outcrop of the Unicoi and extend to the base of that formation but do not penetrate it, indicating an igneous unconformity at the base of the Unicoi formation.

From the evidence cited in Clarke and Loudoun counties and in the Elk Creek anticline, Va., it is established that the late pre-Cambrian deformation was not marked by orogeny in these areas but by broad epeirogenic uplifts of such an extent that an erosional break exists between the late pre-Cambrian and Lower Cambrian rocks. The Swift Run tuff (the equivalent of the Lynchburg gneiss), Catoctin basalt, and the Mt. Rogers volcanic series, therefore, are not part of the Lower Cambrian sequence.

In central Virginia it has been claimed (Bloomer and Bloomer, 1947, p. 97, 103–106) that the Swift Run tuff (Oronoco formation), Catoctin basalt, and Lynchburg gneiss are Lower Cambrian. The Montebello thrust block (Fig. 6), containing the Swift Run tuff (Oronoco formation) and Catoctin basalt, is thrust westward over the Buena Vista anticline and the Fullhardt Knob block, in which the Catoctin basalt and the Swift Run tuff are absent, and the Lower Cambrian quartzites having thin flows, tuffs, and arkose at the base directly overlie the injection complex. Because the Lower Cambrian facies in the Fullhardt Knob block is like that in the Buena Vista anticline (Table 5, column a), the writers suggest that the fault which bounds this block, here called the Fullhardt Knob thrust, originated as a break in the eastern part of the anticline. In the vicinity of Roanoke, Va., this thrust has been called the Blue Ridge thrust (Woodward, 1932, p. 80–83, Pls. 1, 2). Bloomer and Bloomer (1947, p. 97–102) state that in central Virginia the Catoctin is a hydrothermally altered andesite and that "The Oronoco thins northwestward until the Catoctin rests directly on the pre-Cambrian crystalline complex." They evidently regard the thin flows and tuffs at the base of the Lower Cambrian in the Buena Vista anticline (Table

5, column a; Fig. 6) as the equivalents of the Catoctin basalt and the underlying Swift Run tuff (Oronoco formation). The Catoctin basalt is in general a massive, dark-green, hydrothermally altered olivine basalt with ophitic texture (Williams, 1892, p. 490–492), and contains andesite flows of lesser extent. Its thickness has been estimated to be 1000–2500 feet. It does not resemble the thin flows, tuffs, and arkose at the base of the Cambrian and, as has been shown, is separated from these basal Cambrian beds elsewhere by an erosional break (Fig. 8). This stratigraphic evidence does not support the conclusion (Bloomer and Bloomer, 1947, p. 106) that there is "an overlapping relationship between the Oronoco, Catoctin, and Unicoi formations" and the resultant inference that the Oronoco formation (Swift Run tuff), Catoctin basalt, and Lynchburg gneiss are Lower Cambrian.

Some geologists (Keith, 1895a, Pl. 22, p. 314–318; Furcron, 1934, p. 406–407; Cloos and Hietanen, 1941, p. 80, 83, Fig. 29) have claimed that granites, which the writers regard as part of the injection complex, intrude the Catoctin basalt. The writers have not seen any outcrop where this relation can be demonstrated. They have given evidence (Stose and Stose, 1946a, p. 27–28) showing that the Catoctin basalt and Swift Run tuff are not penetrated by granite and were deposited on a floor composed of granite and metasediments which comprise the injection complex. This evidence includes the facts that diabase dikes (Fig. 8), which intrude the injection complex and are related to the Catoctin basalt, have chilled borders and cut across primary layering of the granite gneisses; that the Swift Run tuff contains pebbles of the underlying granitic rocks; and that the volcanic rocks are much less metamorphosed than the injection complex.

The writers conclude, therefore, that the Swift Run tuff and equivalent Lynchburg gneiss, the Catoctin basalt, and the Mt. Rogers volcanic series were extruded or deposited on the injection complex, are separated from the conformable series of Lower Cambrian quartzose rocks by an erosional unconformity, and are of late pre-Cambrian age.

COMPARISON OF OCOEE SERIES WITH LYNCHBURG GNEISS

The Lynchburg gneiss is a biotite-muscovite gneiss and mica schist, in places garnetiferous, which was derived from the metamorphism of arkosic quartzite and interbedded black graphitic slate. It contains conglomerate at the base. The Ocoee series is composed in part of similar sedimentary rocks. The Lynchburg gneiss throughout its extent in Virginia contains intercalated flows and basic sills related to the overlying Catoctin basalt. In North Carolina in the areas of Lynchburg gneiss north of Boone and at Spruce Pine (Fig. 6), the Catoctin basalt is not present above the gneiss, but the gneiss contains similar basic flows and related sills. Such igneous rocks however are almost entirely absent in the Ocoee series. In North Carolina and Tennessee the Ocoee series overlies the west side of a wide area of the injection complex, while the Lynchburg gneiss in the syncline near Spruce Pine (Fig. 6) overlies the east side, and the quartzose conglomerate at the base of the Lynchburg in that syncline is only 28 miles southeast of the base of the Ocoee series. It is probable therefore that the Ocoee series and the Lynchburg gneiss were deposited in different parts of the same basin in late pre-Cambrian time and that vulcanism was restricted chiefly to the eastern part of the basin.

STRATIGRAPHIC RELATIONS

The writers have found Lower Cambrian quartzite of the Unicoi formation in stratigraphic relation to the Ocoee series at one place—namely, in the gorge of Nolichucky River in eastern Tennessee, $1\frac{3}{4}$ miles southeast of Unaka Springs. In good exposures on the Carolina, Clinchfield, and Ohio Railroad, where the railroad follows the south side of the river, the well-bedded quartzites of the Unicoi formation dip 60°–70° NW., and the arkose of the lower part of the formation, beneath the medial basalt flow, has at its base a 15-foot bed of conglomerate composed of densely packed, round, white quartz pebbles, typical of basal beds of the formation elsewhere (Pl. 4, Fig. 1). This conglomerate overlies rocks to the southeast whose lithology is like that of the Ocoee series. The outcrops of the conglomerate are separated from those of the underlying rocks by a narrow ravine; since the rocks exposed on either side of the ravine show no disturbance such as would be produced by faulting, it is assumed that contact of the conglomerate with the rocks beneath is depositional. The rocks beneath the conglomerate are poorly bedded, white to gray quartzite containing intercalated dark banded graywacke and apparently belong in the lower part of the Great Smoky quartzite. To the southeast still lower beds, consisting of ferruginous arkosic quartzite and dark-gray to black graywacke, probably Hurricane, dip west off the granite gneiss of the injection complex exposed at Lost Cove station. The bedding in the Unicoi formation and in the conglomerate at its base is parallel to that in the underlying rocks here identified as Great Smoky quartzite,and both groups dip northwest. There is no evidence of a structural unconformity between them, but the presence of the conglomerate at this contact suggests a disconformity such as the writers have shown to exist between the Lower Cambrian and late pre-Cambrian rocks in northern (Fig. 8) and southern Virginia (Fig. 10).

The writers' interpretation (Fig. 6) is that the Great Smoky overthrust originated on the southeast side of the Elk Creek anticline and carried the southeastern part of the Holston Mountain block over the northwestern part. As the Unicoi formation is characteristic of the Lower Cambrian facies in the Holston Mountain block, the same Lower Cambrian facies should occur in the Great Smoky block. This facies is in that block in Buffalo Mountain (Fig. 6) which contains the Unicoi formation with its characteristic medial flows. The probable disconformity at the near-contact of the Unicoi formation and the Ocoee series in the gorge of Nolichucky River and the structural interpretation that the Unicoi formation should and does occur elsewhere in the Great Smoky block make it more certain that the Unicoi overlies the Ocoee series with a depositional contact.

The writers base their conclusion that the Ocoee series is of late pre-Cambrian age on the following evidence: The series is younger than the early pre-Cambrian injection complex because it overlies the complex with a pronounced stratigraphic, structural, and metamorphic break; the Ocoee series does not resemble Lower Cambrian quartzose rocks in lithology or manner of deposition; the Ocoee resembles the Lynchburg gneiss of late pre-Cambrian age and was probably deposited in the same basin; Lower Cambrian quartzites apparently overlie the Ocoee series with an erosional disconformity as a result of late pre-Cambrian epeirogenic movement.

FIGURE 11.—*Diagrammatic section of the gradation of Ordovician sediments*

Showing gradation of calcareous sediments southeastward into clastic shore facies, comprising an expanded Tellico formation. Osvs, Ot, Oas, Oks, sandstone beds in Sevier, Tellico, Athens, and Knox. Otx, expanded Tellico formation.

CONCLUSIONS

The Ocoee series is made up of dark graphitic graywacke, slate, and quartzite in which the coarse-grained detrital material is largely angular. The formations of the series vary greatly in thickness. Its metamorphism is commonly of low rank and is of high rank in the southeast part of the belt where the series is intruded by pegmatite. The series unconformably overlies the early pre-Cambrian injection complex; it does not resemble Lower Cambrian quartzose rocks and is separated from them by an erosional unconformity—hence it is of late pre-Cambrian age. The series has been thrust northwestward on the Great Smoky thrust fault and on the equivalent Cartersville thrust; in places it has ridden over unmetamorphosed sandstone, shale, limestone, and limestone conglomerate which were formerly included in the Ocoee series; these rocks are named the Tellico formation, a shore facies of marine sediments in large part equivalent to the Middle Orodvician Tellico sandstone. The Tellico formation overlies the Athens shale and Knox dolomite in the Pulaski thrust block. Overthrusting on the Great Smoky thrust has brought the Ocoee series in juxtaposition also with thrust slices and klippen of the Chilhowee group which contains Lower Cambrian quartzose facies characteristic of the Pulaski block. These slices were probably broken from anticlines in that block and were pushed in front of the northwestward-moving Great Smoky block. The first folding and metamorphism of the Ocoee series may be of Taconic or late Paleozoic age. The overthrusting and accompanying second deformation of the series occurred during Appalachian orogeny because the overridden blocks contain Carboniferous rocks.

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Deadwood Creek
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Homestake Open Cut
DeSmet syncline
Independence anticline
Lead syncline
Poorman Gulch
Gold Run
Pierce anticline
Whitewood Creek
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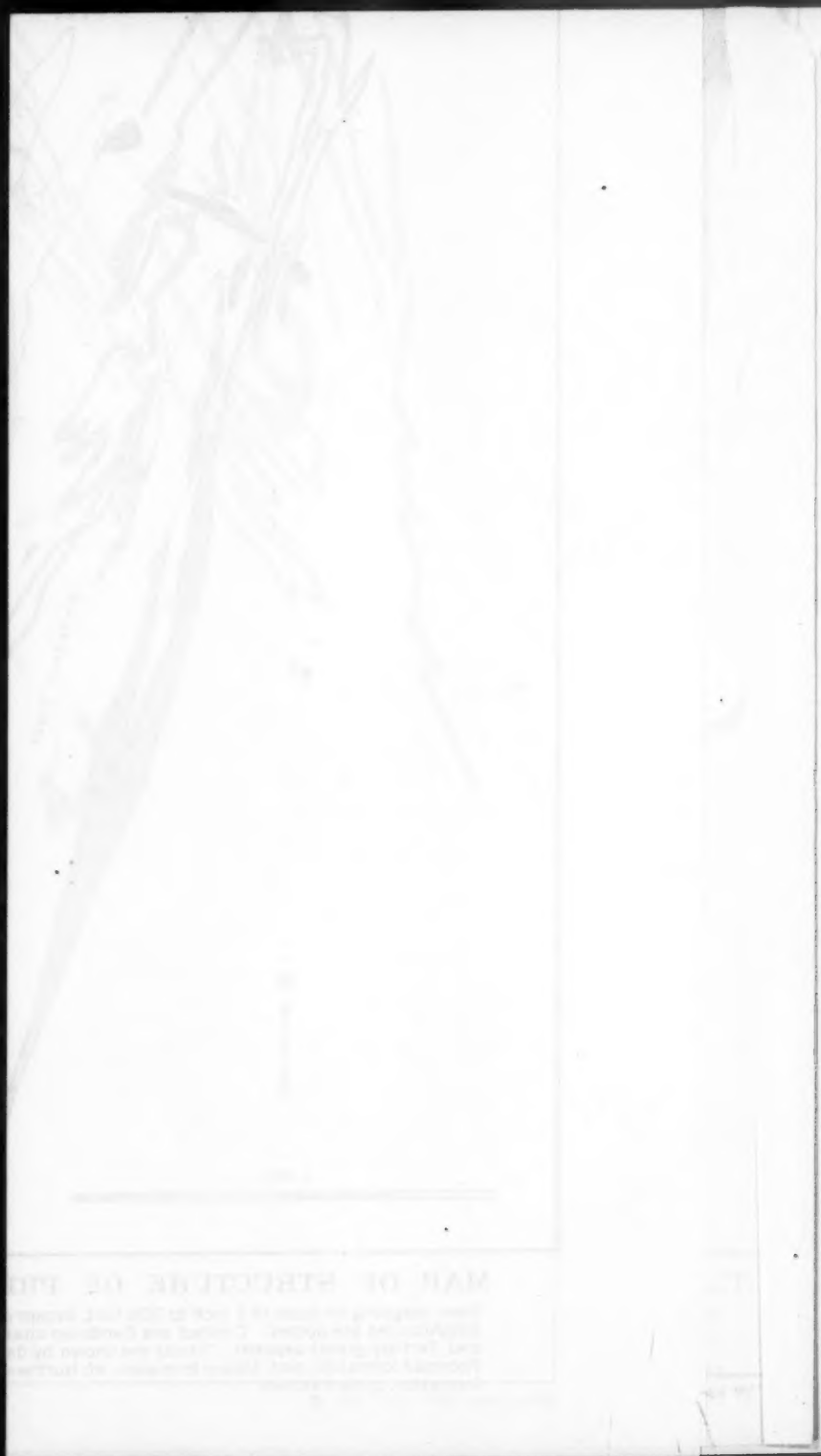
MAP OF STRUCTURE OF PRE-CAMBRIAN ROCKS IN THE LEAD DISTRICT.

From mapping on scale of 1 inch to 200 feet, except a small part on scale of 1 inch to 1000 feet. Tertiary intrusives are in solid black. Amphibolites are dotted. Omitted are Cambrian strata, small amphibolites, small Tertiary intrusives (including sills in the Cambrian), and Tertiary gravel deposits. Faults are shown by dashed lines.
Poorman formation, pmf; Ellison formation, ef; Northwestern formation, nwf; Flag Rock formation, frf; Grizzly formation, gf; Homestake formation, cross hatched.



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STRUCTURE OF A PART OF THE NORTHERN BLACK HILLS AND THE HOMESTAKE MINE, LEAD, SOUTH DAKOTA

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ABSTRACT

The pre-Cambrian rocks of the northern Black Hills in the vicinity of the Homestake Mine show the effects of extreme isoclinal folding on which has been superimposed shear folding crossing the isoclinal axial planes at small angles. An attempt has been made to distinguish the two types of folding in all scales of folds from those of microscopic size to those large enough to appear on the regional map.

Enough Cambrian cover remains in this area to permit contouring the top of the pre-Cambrian rocks. These contours, which essentially measure post-Cretaceous

deformation, show a close correlation between doming of the district and the intrusion of a stock of Eocene age. Several other domes in the northern Black Hills, previously called laccoliths, are believed to be due to stock intrusion.

The deformation of pre-Cambrian rocks by the Tertiary intrusive bodies is more difficult to interpret. There are, however, many small-scale deformations attributable to igneous intrusion, and probably there are large ones as well. Whether the shear folds crossing the isoclinal axial planes are due to Tertiary igneous intrusion or, like the isoclinal folds, are of pre-Cambrian age cannot be answered at this time.

INTRODUCTION

The northern Black Hills of South Dakota in the vicinity of the Homestake Mine offer what is believed to be an opportunity unequalled elsewhere for a study of complex structures in highly metamorphosed rocks. The pre-Cambrian rocks are for the most part sedimentary rocks with well-defined bedding planes and a number of good horizon markers; very valuable ore deposits are confined to one stratigraphic horizon and are so definitely localized by structure that it is imperative to study every detail of the structure; and erosion has exposed large areas of the pre-Cambrian rocks but has left enough of the Cambrian rocks to permit reconstructing the shape of the deformed base of the Cambrian and comparing it with the structure of the pre-Cambrian rocks.

This paper is mainly the work of the three authors, but it is based in part on work of the earlier members of the Homestake Geological Department: J. O. Hosted, L. B. Wright, A. B. Yates, and C. N. Kravig. We have also been assisted by the work of a number of past or present associates: T. A. Dodge, R. G. Wayland, G. C. Mathisrud, T. M. Rizzi, and D. C. Lavier. D. T. Griggs carried on a special study of the mechanism of dike intrusion. An early version of the manuscript was read critically by Professors E. S. Larsen, Jr., Kirk Bryan, and M. P. Billings of Harvard University, Professors G. M. Schwartz and J. W. Gruner of the University of Minnesota, D. H. McLaughlin, president of the Homestake Mining Company, D. T. Griggs, and T. A. Dodge, to all of whom we are greatly indebted for much general and specific criticism and discussion. Finally, visitors too numerous to mention have contributed directly or indirectly by discussion and searching questioning.

Brief mention of the structure of the Black Hills pre-Cambrian rocks is contained in papers by Newton and Jenny (1880), Van Hise (1890), and Jaggar (1901). The results of the first comprehensive study of the structure in the vicinity of the Homestake Mine were published by Paige (1913a), and those of the first detailed studies of the structure exposed in the mine workings were published by Hosted and Wright (1923) and by Paige (1924) and Darton and Paige (1925). Further results of structural studies in the Homestake Mine have been published by Gustafson (1930; 1933), Yates (1931), and McLaughlin (1931). Papers which have discussed the pre-Cambrian structure in different parts of the Black Hills incidental to other studies are those by Balk (1931), Harder (1934), Runner (1934), and Dodge (1935; 1942).

STRUCTURE OF THE PRE-CAMBRIAN ROCKS

STRUCTURAL MARKERS

Detailed descriptions of the structure of highly metamorphosed rocks are more understandable if preceded by a discussion of the horizon markers by which the

structure is revealed. In the Lead district the broad structural relations are shown by the outcrop patterns of the sedimentary formations. Those formations have been separated from each other wholly on lithologic evidence, but, within the area of this district, the continuous lithologic units probably are essentially contemporaneous. The problem of determining the major structure of the Lead district has been, however, more difficult than a mere mapping of rock types. Until the process of metamorphism was understood and its products recognized, structural studies were seriously hampered. Even more serious in this area of intense isoclinal folding was the difficulty of correctly determining the stratigraphic succession. The succession, described in detail in an earlier paper (Noble and Harder, 1948) is as follows:

Tertiary, Quaternary, and Recent gravel deposits
(at least 90 feet; probably much more)
~~~~unconformity~~~~
Tertiary intrusive rocks
Cambrian Deadwood formation (300 to 500 feet)
~~~~unconformity~~~~
Pre-Cambrian intrusive rocks
Pre-Cambrian sedimentary rocks
Grizzly formation (possibly 3000 feet or more)
Flag Rock formation (possibly 5000 feet)
~~~~unconformity~~~~
Northwestern formation (possibly 4000 feet)
Ellison formation (3000 to 5000 feet)
Homestake formation (200 to 300 feet)
Poorman formation (possibly 2000 feet or more)

This tabulation is the result of continuous study by a number of geologists over a great many years, during which time an understanding of the stratigraphy and of the structure have necessarily progressed together. The real difficulty in determining the correct stratigraphic succession has been the recurrence of identical rock types in different formations. For example, cummingtonite schist, an unusual rock, occurs not only in the Homestake formation but also at five or six other horizons in the Lead district. It is believed that the stratigraphic column is now correct in all essential features; but many changes have been made within the last few years, and it may be that minor changes will be made in the future. It does not seem probable, however, that any future changes in the stratigraphy will necessitate a major revision of the structure.

Two contacts between formations have proved to be of outstanding value as horizon markers: that between the Homestake and Ellison formations; and that at the base of the Flag Rock formation. It is most fortunate that the Homestake formation, a narrow unit of unusual character underlain and overlain by formations which can be distinguished with reasonable certainty, is the only one susceptible to extensive replacement by ore solutions and hence is the one most thoroughly explored in underground and surface prospecting.

The minor structural features are disclosed by bedding planes still preserved in spite of the intense folding and metamorphism. Bedding planes in this district are characterized by one or more of the following features: (1) change of lithologic character, as from a phyllite to a quartzite (the pods and bands of recrystallized chert
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are a special type of lithologic change); (2) graphitic banding; and (3) color banding. The first is clearly a primary sedimentary feature, and the second and third, everywhere parallel to the first, must be related to bedding although they may be determined by metamorphic minerals. In the mine workings, detailed mapping of the attitudes of the bedding planes and the minor folds shows the main structural features; but on the surface, where exposures are not abundant, the bedding planes are of little value for structural study, and the tracing of lithologic units gives the best information.

Still another minor structural feature of considerable aid in the mine mapping is the parallelism of cleavage and bedding around the crests of folds. Throughout most of the mine area this strict parallelism of cleavage and bedding causes small folds to break out intact on the walls and backs of workings, leaving the structure conspicuously recorded. The manner in which cleavage and bedding may become parallel around the crest of a fold will be discussed later.

Other features which have been used in structural studies in other areas are minor folds, the relation of cleavage to bedding, and the attitudes of primary structural features such as cross bedding, ripple marks, and gradational bedding. Minor folds (sometimes called drag folds) are very abundant in the rocks of the Lead district, and under some conditions they are useful in locating the positions of major axial planes. Under conditions of cross folding, however, the minor folds do not point out the major axial planes and are of little aid in the recognition of the larger isoclinal folds. The intersection of bedding by cleavage, used in some places to distinguish overturned beds from those in normal position, is of no significance in the Lead district. In a few places cleavage intersects bedding in the arches of folds in the theoretical manner, but it becomes parallel to the bedding at so short a distance from the axes that the relations are of no use in recognizing overturned limbs of folds. However, most folds in the vicinity of the Homestake mine either show cleavage parallel to bedding around the fold or show no good cleavage on the crest of the fold. Cross bedding, ripple marks, and gradational bedding do not occur in the pre-Cambrian rocks of this district. Either they were never present or they have been obliterated by the metamorphism.

BROAD STRUCTURAL FEATURES

The dominant structure of the pre-Cambrian rocks of the Lead district is a complex anticline. The axial plane strikes a little west of north and dips east, and the axial line plunges at about 25° S.E. This anticline—the Poorman anticline—extends from the northwest edge of the map area through Poorman Gulch and the west edge of Lead to a point on the south edge of the map area (Pl. 1). The anticline is tightly closed to a remarkable degree. The outcrop of the Poorman formation in the axis of the plunging anticline on the northwest edge of the map is 1200 feet wide, and at a point 2½ miles to the south it is just as wide although it narrows farther south. The Lead syncline is a somewhat broader structure with its axis about 2500 feet east of the axis of the Poorman anticline. Many very narrow minor anticlines and synclines occur on the limb between the Poorman anticline and the Lead syncline and in the bottom

of the syncline. East of the Lead syncline occurs a group of folds comprising two very narrow anticlines and a narrow syncline. The syncline is called the De Smet syncline, and the Pierce anticline and Independence anticline are east and west of it respectively. These structures crop out in the Homestake open cut and contain most of the mineable ore of the district. They are, however, minor structural features in comparison with the Poorman anticline. East of the Pierce anticline are many minor anticlines and synclines, but, except on the scale of the mine maps, most are inconspicuous features. Near the northeast edge of the map area an anticline extends beyond the limits of the map and may be of considerable size. West of the Poorman anticline are several tightly compressed anticlines and synclines. The axial planes of all the folds west of the Pierce anticline strike a little west of north, whereas those of folds east of the Pierce anticline strike about north or a little east of north. All axial planes and limbs of folds dip about 65° E., and all folds plunge to the southeast. The plunges are as low as 10° or 15° in the southwest part of the district, except for local differences, but they gradually increase to about 45° in the northeast part of the district.

The broad structural relations of the region are those of minor folds related to the dominant Poorman anticline, but many structural features are not in accord with this relationship. One discordant feature is the change in strikes of the axial planes from northwest to northeast that takes place quite abruptly on a line east of the Pierce anticline (Pl. 1). This is not merely an example of the convergence of the axial planes of minor folds toward the axis of an anticlinorium, because the structure east of the Pierce axis is out of accord with that to the west. A second discordant structural feature is that of a flexing of the isoclinal axial planes which will hereafter be called cross folding. Plate 1 shows many cross folds and "reverse drags", *i.e.*, minor folds which are opposite to the theoretical direction. The axial plane of the Poorman anticline, because of a series of cross folds, crowds against and apparently cuts off the axial plane of the Lead syncline and the axial planes of several minor folds.

Although faulting on a major scale is rare in the immediate vicinity of the Homestake Mine, one fault near the eastern edge of the district has been traced throughout the full length of the map area. It strikes a little west of north and probably dips eastward. The displacement, measured on the base of the Deadwood formation, ranges from about 120 feet on the southern edge of the map area to about 600 feet on the northern edge, everywhere with the east side up. The strike of the fault is so near that of the pre-Cambrian rocks that almost no displacement of the pre-Cambrian rocks can be observed. This fault is occupied by a large wedge-shaped stock of porphyry in the northern part of the map area and by smaller dikes at several other places. The large wedge-shaped stock is accompanied by many other smaller faults, some of which have a roughly radial pattern. Another fault, in the vicinity of Whitewood Creek near the southwest edge of the map, has displaced the base of the Deadwood formation about 100 feet and has displaced several limbs of the Homestake formation. A few other faults displace the base of the Cambrian strata a few feet but leave no measureable effects in the pre-Cambrian rocks that can be distinguished from the isoclinal or cross folding.

DETAILS OF STRUCTURE

The outcrops of the isoclinal folds are extremely long and narrow, and the tight folding of the Poorman anticline is duplicated by many minor folds. Clearly the elongation passed far beyond that attributable to competent folding and must have been accomplished by slipping along closely spaced shear planes that were about parallel to the axial planes. The crests of many isoclinal folds are split into a number of folds of approximately equal size, not true minor folds. Good examples are the Independence anticline and Lead syncline and others on the mine maps. A similar splitting of the axes of folds has been described by Balk (1936, p. 723).

The cross folding, like the isoclinal folding, must have been produced by slipping along closely spaced shear planes, and the two types of folding cannot always be distinguished. The most convincing evidence that isoclinal folds have been refolded is afforded by the deformation of closed folds like the Pierce anticline and the De Smet syncline and by some minor folds (Figs. 1, 2). But, if horizon markers on opposite limbs of the isoclinal fold are not fairly close together, the cross folding is not so easily recognized, because only one limb of the isoclinal structure is folded into abnormal shapes or "reverse drags" whereas the other limb is folded into minor folds which appear to be in accord with the isoclinal folding. In Figure 2, showing a cross fold in the De Smet syncline, the northeast limb of the syncline appears to be folded into minor folds in accord with the isoclinal folding, whereas the southwest limb is clearly cross folded. The "reverse drags" plunge to the southeast nearly parallel to the plunge of the isoclinal folding. In this manner they produce in part of the fold a structure which is difficult to name because of the inverted position of the strata. These upside-down structures are not folds which have been rotated through 180 degrees but are newly formed folds. Harder (1934) has attributed a large upside-down structure at Rochford to large-scale cross folding.

As a result of cross folding, very narrow isoclinal folds may become so tightly closed that some formations are separated into pods or lenses. In a horizontal plane, such as a mine level, the pods appear to be wholly detached from the rest of the formation; but they are probably all attached either above or below and can be likened in shape to stalactites and stalagmites, with central axes inclined to the southeast a little more steeply than the plunge of the regional structure. The relations of isoclinal folding and cross folding within the main mine area cause the synclinal pods to become detached and to end downward whereas anticlinal pods become detached and end upward.

Figure 3 shows the principal cross folds identified in the main part of the Homestake mine.

Two types of cross folding occur within the Lead district, distinguished by the orientation and direction of displacement. Within the mine area the displacement in plan is similar to the shape of a reversed letter "N" (Figs. 1–3). Shearing on planes striking about north and dipping steeply east probably causes this displacement. Because the displacements are deduced wholly from the shapes of maps of plunging structures, it is not possible to determine whether the displacements are wholly horizontal or wholly vertical or partly horizontal and partly vertical. A displacement of the west side up will give an apparent displacement like that shown in

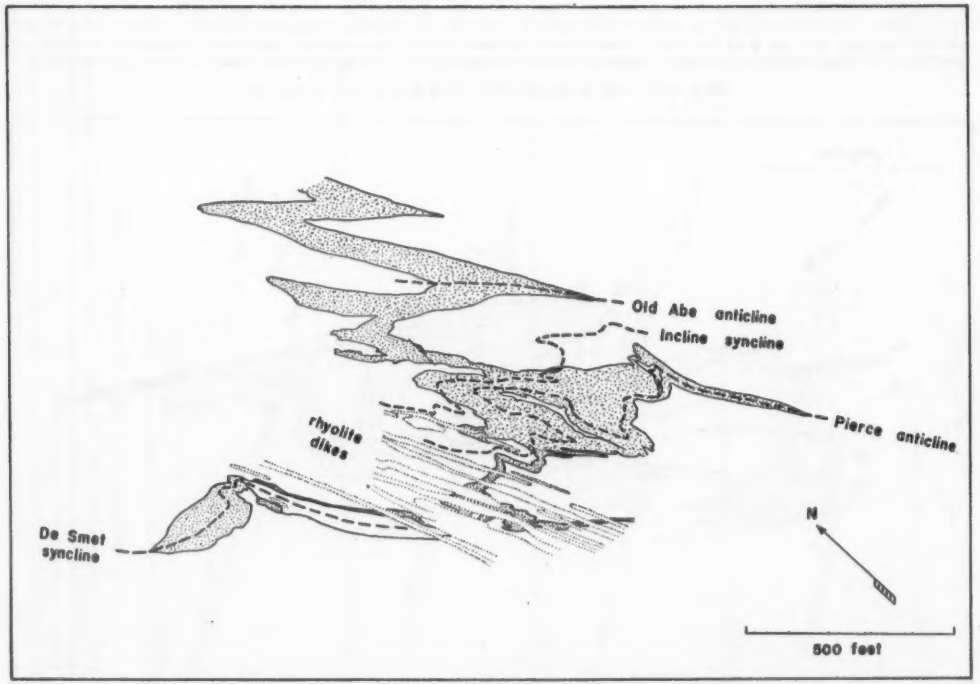


FIGURE 1.—*Cross-folded axial planes on the 2600 level, Homestake Mine*

Traces of axial planes of the Old Abe anticline, Incline syncline, Pierce anticline, and De Smet syncline shown by dashed lines. Their positions are known from mapping of stopes and workings on this level and on adjacent levels. All axial planes except that of the Old Abe anticline have been cross folded.

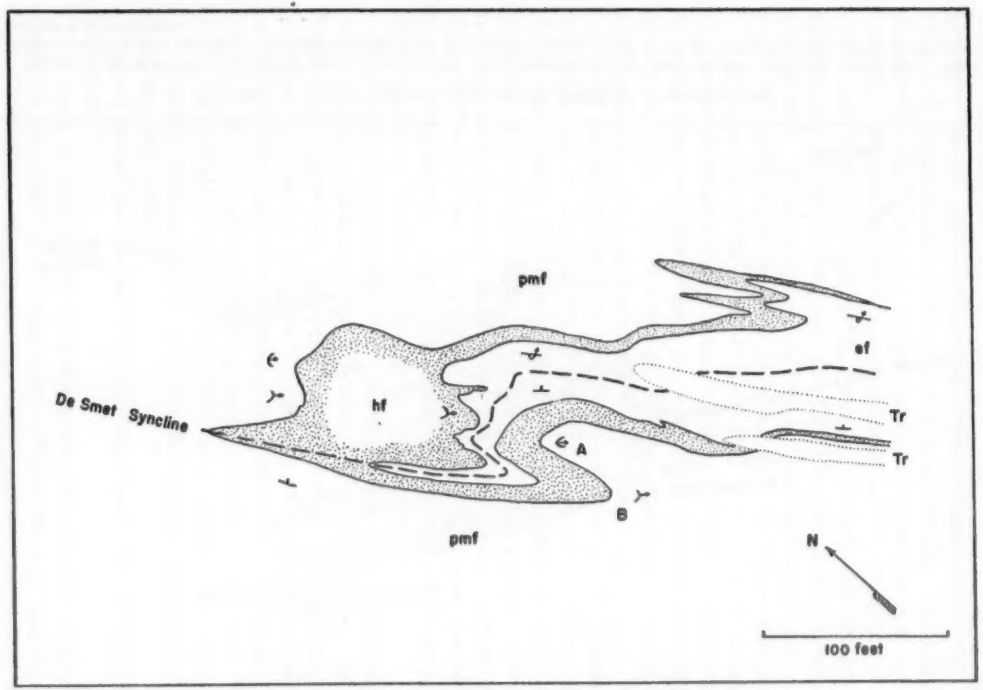


FIGURE 2.—*Map of a part of the 2300 level, Homestake Mine*

Axial plane of the De Smet syncline shown by dashed line. All parts of the structure plunge southeast (to right on sketch). At *A* the rocks have the shape of a syncline, but the oldest formation (the Poorman formation) is in the trough. Similarly at *B*, the rocks have the shape of an anticline, but the youngest formation (the Ellison formation) is in the axis. Across the axial plane of the structure, on the other hand, the minor folds are not in inverted position, although they also are probably due to cross folding. Ellison formation, ef; Homestake formation, hf; Poorman formation, pmf.

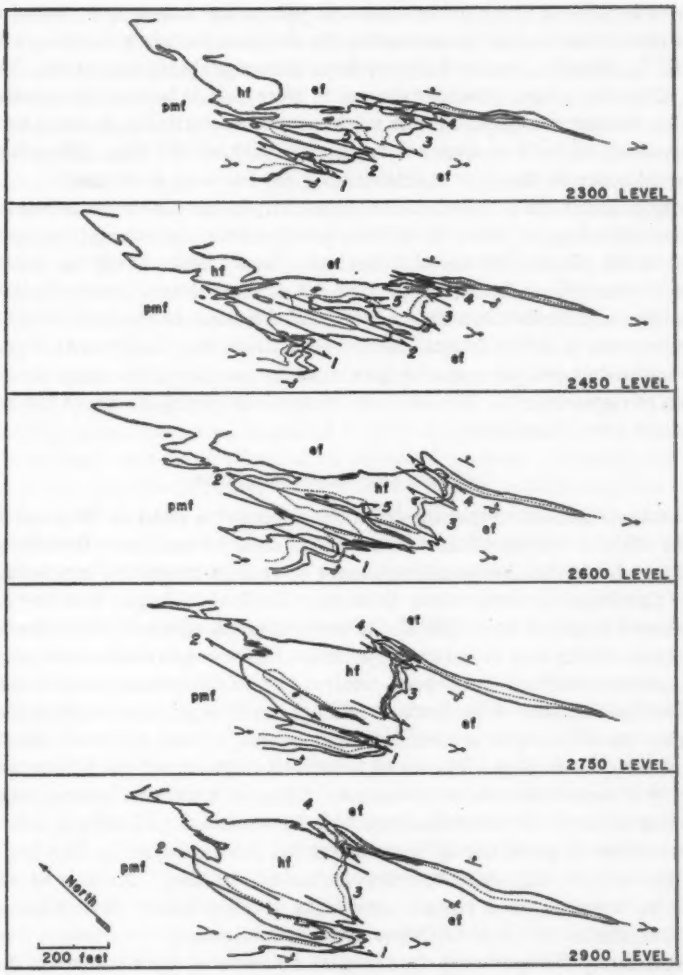


FIGURE 3.—*Structure of a part of the Homestake Mine*

Axial plane of Pierce anticline shown by dotted line. Inferred positions of cross folds in the Pierce axis shown by shear couples and numbers 1–5. Inasmuch as the displacements on the cross folds are inferred from the shapes of plunging structures, there are no criteria to determine whether the displacements are horizontal, as shown, or vertical. If displacements are vertical, relative movement is west side up for cross folds 1–4, and east side up for cross fold 5. Ellison formation, ef; Homestake formation, hf; Poorman formation, pmf.

plan. Tiny closely spaced shear planes parallel to the direction of the shear couple are visible in some cross folds, but in many instances the flexing of the isoclinal axial planes and the creation of new folds, some inverted, are the only evidences of cross folding. The new folds produced by this type of cross folding plunge 40°–45° S., a little steeper than the plunge of the isoclinal folds. In several other areas, however,

notably in Deadwood Creek in the northwest part of the map area of Plate 1 and in the Caledonia structure of the mine area, the displacements in plan make a normal letter "N". Shearing on northwest-striking planes probably causes this displacement. The shear planes probably dip east. Here, too, it has not been possible to determine whether the displacements are horizontal or vertical. A displacement of the east side up will give an apparent displacement in plan like this. The attitude of the new folds due to this type of cross folding has not been determined.

No major faults have been detected with certainty in the mine area, but some zones of intense stretching of limbs of folds in pre-Cambrian rocks may correspond to mapped faults of the Deadwood formation. Small faults, however, have been mapped at many places in the mine. On a few of these the maximum displacement is probably about 50 feet, in a nearly horizontal direction, but on most of the faults the displacement is only 1–2 feet. Almost all faults strike a little west of north and dip steeply either east or west. Adjacent faults may have the same or opposite direction of displacement. The attitudes of the fault planes are about the same as those of one set of cross folds.

TEXTURES AND MICROSTRUCTURES

The textures and microstructures of the pre-Cambrian rocks of the Lead district show the effect of the complicated course of deformation and throw further light on its nature. Logically, the specimens should be studied in order of increasing complexity, starting with those which show only isoclinal folding. This method has been followed in spite of some difficulty in recognizing the effects of later deformations due to cross folding and to igneous injection. In some areas where structures are simple and cross folding has not been detected part of the rock probably still retains the undeformed texture of the isoclinal folding; but in other areas, such as the main mine area, cross folding is so abundant that probably every specimen shows some effects of the cross folding. Moreover, some rocks of the Lead district are unsuited to a study of the relations between textures and major structures because they show no bedding either in the hand specimen or on the outcrop. Finally, in some areas notable increase of grain size by recrystallization has destroyed the bedding planes and so prevents a comparison between texture and bedding. In spite of all these limitations, however, there remain some areas which probably show only isoclinal folding undisturbed by cross folding or by igneous injection.

Approximately 100 specimens were selected for study of those textures and microstructures assumed to be produced only by isoclinal folding. These specimens were selected by elimination of unsuitable material on the basis of the limitations just given. Of the total list, fully half of the specimens show cleavage strictly parallel to bedding. Cleavage in these specimens is marked by the orientation of biotite and muscovite and the dimensional orientation of elongated quartz grains. There is in addition a little random biotite in some specimens. The optical orientation of quartz has not been studied. A few specimens show definite elongation of clastic grains of quartz in the plane of the cleavage and the bedding, and one specimen in which clastic grains of quartz are abundant and moderately coarse shows a definite elongation of the grains in the direction of the tectonic axis or plunge. Specimens

from the lower grades of metamorphism show more nearly perfect orientation of micas than those of coarser grain size from the zones of higher metamorphic grade. Most of these specimens, perhaps all, come from the limbs of isoclinal folds. In addition, one specimen from the crest of a small fold shows cleavage strictly parallel to bedding although the degree of orientation of the micas is not high.

A few specimens of isoclinally folded rocks show definite shear planes crossing the bedding and the cleavage, which parallels bedding. Some of the shear planes have caused no reorientation of micas, but some have changed the orientations in one way or another. In some specimens new or reoriented micas are now oriented in the plane of shearing. In others, especially those in which the shearing crosses bedding at a high angle and in a zone up to 1–2 mm. wide, the micas within the shear zone bisect the acute angle between the shear zone and the main cleavage, either due to a bending of the original grains or to a new crystallization (Pl. 2).

Where the micas are bent by a shear zone, the new positions usually show the direction of relative displacement on the shear plane. In all specimens of this group, the bedding planes have been displaced as if by a normal fault—that is, the rocks have been lengthened in the direction of the bedding plane. Probably every stage can be found from that of planes which cross the bedding and cleavage but cause no reorientation of micas to a final stage in which all earlier orientation of micas has been obliterated by reorientation parallel to the shear planes. A few specimens show what is probably almost the final stage in this process; lenses of unsheared rock preserve relics of an earlier orientation probably parallel to bedding, whereas the rest of the section shows a new cleavage crossing bedding. In hand specimens and outcrops the wider shear zones with bent micas make small crenulations, whereas the narrower shear planes make either a true cleavage or a "false cleavage" (*Ausweichungsclivage*) crossing bedding. Similar occurrences have been described by Dale (1899), Behre (1926), Knopf (1935), Cloos (1937), Chapman (1939), and others.

Specimens of another group of isoclinally folded rocks show in addition to a good cleavage parallel to bedding a second less-pronounced cleavage inclined to the first one and to the bedding. The second cleavage in these specimens is a true cleavage in the sense that it results not from closely spaced shear planes but from a parallel orientation of micas. In some of the specimens both cleavages are produced by the same kind and shape of muscovite or biotite. But in a few specimens the second cleavage is produced by parallel crystallographic orientation of stubby crystals of biotite, whereas the better cleavage is produced by orientation of thin plates of muscovite or biotite. The plane of orientation of the stubby crystals intersects that of the thin plates at high angles in some specimens, at low angles in others. One specimen from near the crest of a fold showed three kinds of cleavage: a good cleavage parallel to bedding produced by orientation of plates of biotite; a fairly good cleavage crossing the bedding at angles of 15°–20°, produced by orientation of plates of biotite; and a poor cleavage at right angles to the second one, produced by crystallographic but not dimensional orientation of stubby grains of biotite.

Still another group of specimens of isoclinal folding shows no cleavage parallel to bedding, but cleavage or shear planes, or both, cross bedding. In some specimens the angle between bedding and cleavage is only 5° or 6°, but in other specimens from the

crests of folds the angle is nearly 90°. Some specimens show not only cleavage intersecting bedding, but also shear planes intersecting both cleavage and bedding, with a reorientation of some mica in the shear planes. In all specimens but one, the direction of relative displacement of cleavage by shear planes was in a "normal" direction, in terms of fault displacement; that is, the rock has been lengthened in a direction parallel to the bedding.

It is believed that the order of presentation of these descriptions is the order of creation of the textures and microstructures. In the stage of concentric folding, not now recorded in the Lead rocks, deformation takes place by slipping between bedding planes, and the platy minerals if not already parallel to bedding by deposition are made parallel by rotation and perhaps by recrystallization. During the later period of isoclinal folding, deformation must take place by slipping on shear planes about parallel to the axial planes. Plastic upward movement of deeply buried, structurally isotropic material under the stress of tangential compression may give the force which causes the shearing. On the flanks of the folds the shear planes parallel the bedding planes, but they cross the bedding for short distances on the crests of folds. Where the shear planes cross the bedding, there is either a fracturing (false cleavage) or a local reorientation of micas by bending (crenulations) or by recrystallization (true cleavage). These are intermediate steps in a process of transformation of a cleavage originally parallel to bedding into a cleavage transverse to bedding such as that just described. Relative displacement of the bedding planes by the shear planes can usually be determined in all but the final stages of this transformation of cleavage. The direction of the relative displacement is important. Isoclinal folding by shearing on planes parallel to the axial planes causes a relative displacement of the bedding planes like that of a normal fault, and the rock is extended in the direction of the bedding planes. Cross folding of the sort to be discussed next may cause, on the other hand, a reverse-fault displacement of the bedding planes.

In those specimens which show cross folding or which can definitely be located in zones of cross folding, typically the rock is sliced by closely spaced shear planes, and the direction of relative displacement on the shear planes is in a reversed direction in terms of fault displacement (Pl. 3). Most sections show parallel zones or lenses of undeformed biotite, muscovite, and quartz separated by zones in which the micas are sharply bent and the quartz is reoriented. Some specimens of crenulated schist from zones of cross folding show a distinct localization of the lines of crenulation by garnets, which must mean that some of the cross folding is later than the formation of the garnets. In the undeformed zones the minerals in most specimens are elongated parallel to bedding, but in some they are elongated oblique to bedding. Recrystallization of sericite in the deformed zones is lacking in some sections but is well advanced in others. Where recrystallization is complete or nearly so, the direction of relative displacement usually cannot be determined. A few sections show clearly that in the highly sheared rocks biotite has been largely replaced by sericite. Two sections in which sericite is abundant in the shear planes have some large unsheared plates of prochlorite, in part parallel to the shear planes and in part in random orientation. Possibly this prochlorite has formed as a result of the freeing of magnesia during the replacement of biotite by sericite.

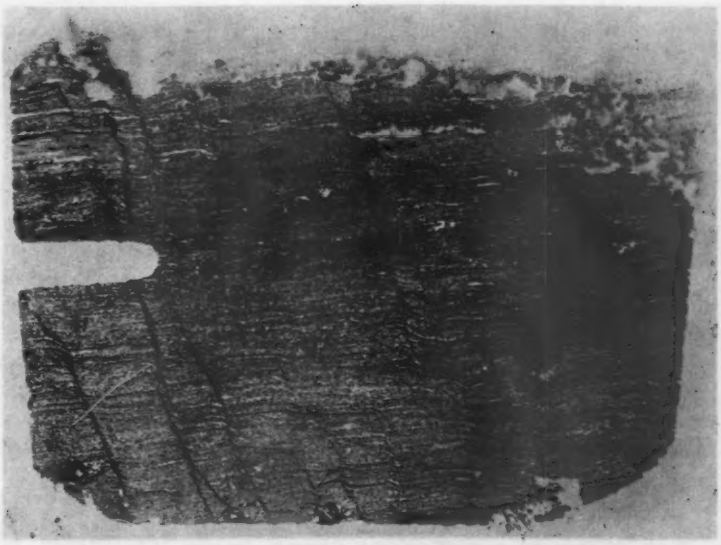


FIGURE 1. THIN SECTION OF CRENULATED MICA SCHIST, PLANE-PARALLEL LIGHT
Biotite is in nearly perfect orientation parallel to bedding, and quartz has good dimensional orientation in same direction. Where small discontinuous planes of shearing or crenulation cross bedding, however, minerals are bent, broken, and in part reoriented. In terms of fault displacement, these shear planes are tiny normal faults, and the rock has been lengthened in direction of bedding. X4 (Compare Plate 3.)

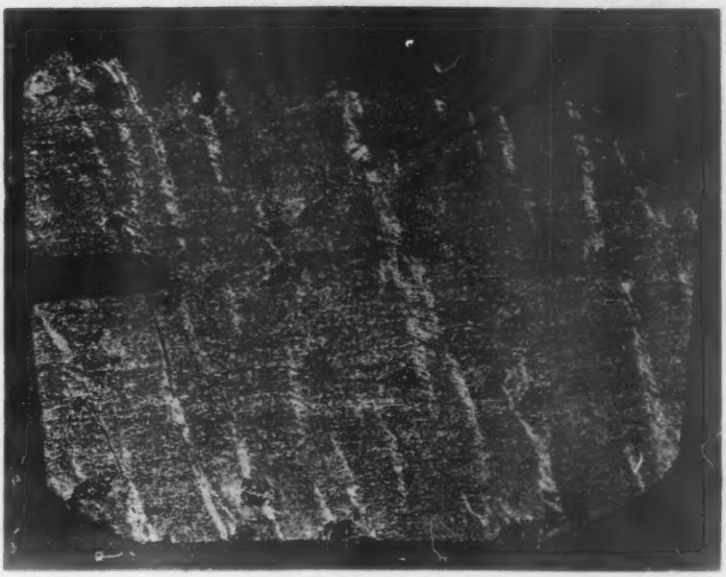


FIGURE 2. SAME SECTION AS FIGURE 1, UNDER CROSSED NICOLS

FOLDED CLEAVAGE

FOLDED CLEAVAGE

Thin section of mica schist from zone of cross folding. Plane-polarized light. Bedding, shown by graphitic and mineral banding, is about vertical in the view. Planes of shearing cross bedding at low angle and dip steeply to right of view. Micas are parallel to bedding between shear planes but are strongly bent in shear planes. Bent micas tend to bisect obtuse angle between bedding and shear planes, and bedding planes are displaced in "reverse" direction, in terms of fault displacement. Rock is therefore shortened in direction of bedding. X4. (Compare Plate 2.)

The shear planes visible in the thin sections of cross-folded rocks are the planes of displacement by which the cross folding has been produced. The total effect of displacement on countless microscopic shear planes is the formation of one of the cross folds previously described. In general, the orientations of the microscopic shear planes are parallel to those at one or the other of the two types of cross folding previously discussed, but some small zones show other directions of shearing.

Although deformation by subparallel shear zones which displace the bedding or cleavage in a reverse direction is common in most of the cross-folded rocks of the Lead district, some specimens have been cross-folded but now show no shear planes. In a few specimens from zones where larger structural relations indicate the presence of cross folding, the orientation of minerals is that of the typical isoclinal folding, parallel or oblique to bedding, with no cross folding now visible. Most of the specimens which fail to show typical deformation or reversed shear planes are quartzites or quartz-rich phyllites of the Ellison formation, and these rocks may in places be deformed by cross folding without retaining much evidence of shearing or reorientation of minerals.

Cummingtonite shows many effects of cross folding. In zones of moderate cross folding cummingtonite rosettes remain essentially radial, but the fibers become curved and strongly twinned. Where the cross folding has been more intense, the fibers become more sharply bent, some of them show transverse fracturing, and in some places the cummingtonite becomes very fine-grained. In the hand specimen or on a crosscut wall the cummingtonite in these zones of intense cross folding shows a fair degree of plane-parallel orientation. Where the cross folding has been most intense, cummingtonite has been largely replaced by chlorite, and only a few sheared aggregates remain.

The origin of the cleavage parallel to bedding around the crests of folds in the pre-Cambrian rocks of the Lead district can be attributed in part at least to the mechanism of cross folding, although it has not yet been demonstrated that cleavage parallel to bedding around the crests of folds cannot have other causes as well. The earliest stage of folding now recognizable in the Lead rocks is isoclinal folding of a moderately high degree. The crests of the folds have not been completely sheared out, however. Thus there are a few small areas in the crests of folds in which the original axial-plane cleavage intersects bedding, but throughout most of the isoclinal areas cleavage and bedding are straight and strictly parallel. In zones of cross folding, on the other hand, the isoclinal limbs of the earlier folds become refolded, and in the early stages of cross folding they form broad, open folds, with no reorientation of schist minerals transverse to bedding. This early stage of cross folding is the one most common in the Lead rocks. A few areas have been further deformed, and some of the schist minerals have become reoriented in the direction of the shear planes, oblique to the bedding and the earlier schistosity.

The course of deformation of the pre-Cambrian rocks as shown by the textures and microstructures of the rocks has thus been studied by comparing the microscopic features with the megascopic features and field relations. Opportunity to study the course of deformation from another point of view is afforded by the porphyroblasts and clastic grains in the pre-Cambrian rocks. The thin sections have, therefore,

been restudied to try to detect evidence of motion in the rocks during the metamorphism. The porphyroblasts in the pre-Cambrian rocks of the Lead district in order of abundance are biotite, garnet, sideroplesite, muscovite, prochlorite, and staurolite; but because of its peculiar properties garnet is the most suitable for a study of the course of metamorphism. It has been found best to study and discuss the effects of the clastic grains of quartz along with those of the porphyroblasts of garnet, as a guide to the interpretation of some of the observations.

The most striking microstructure of the garnets and the one most surely explained is the spreading of the bedding and schistosity to wrap around the garnet grains. This effect, the cause of the "lumpy" appearance of most garnetiferous specimens, is observable in almost every thin section. That it is probably due to the force of crystallization of the garnet crystals is scarcely to be doubted where narrow graphitic bands, narrower than the garnet crystals, are bulged out around the crystals (Pl. 7, fig. 1). Similarly, elongated garnet grains which stand with the long axes at high angles to the bedding and schistosity have deformed the planes of bedding and schistosity well beyond deformation attributable to rotation. Figure 2 of Plate 7 shows an elongated garnet grain which, by growing across the schistosity, has notably deformed the planes of schistosity. Individual biotite plates which wrap around the garnets are in most sections not bent or otherwise deformed and so must have formed or recrystallized contemporaneously with or since the growth of the garnets; but there are a few examples of bent plates of biotite and coarse muscovite. It is not possible in most sections to determine whether the smaller plates of sericitic white mica have been deformed although they probably have been bent. Definite evidence of at least moderate rotation is afforded by a slight asymmetry or S-curve of the bedding and schistosity in passing around some of the garnets. Such occurrences are, however, decidedly rare. Moreover, they do not necessarily call for much rotation.

Comparing these external effects of the garnet crystals with the effects of clastic grains of quartz on the surrounding schist shows that very similar microstructures can be formed by a distinctly different course of metamorphism. The clastic grains are also wrapped by the bedding and schistosity although the degree of bending of graphitic bands and lines of mica plates is not great, nor are individual bands of graphitic material split apart as they are by the garnet crystals. There is, furthermore, very little asymmetry or S-form to the bedding and schistosity as it wraps around the quartz grains although clearly there has been rotation of grains and much movement of material around the grains during the whole course of deformation of the rocks. There is one significant point of difference, however, in the degree of orientation of grains. Those clastic grains of quartz that slow elongation are, almost without exception, elongated in the plane of bedding, and some are parallel to the tectonic axis or plunge. Elongated grains of garnet, on the other hand, are in most sections probably random in orientation; at least, there are many grains which are oblique to the plane of bedding and schistosity, and not a few which are nearly perpendicular to it. There are far too many oblique and perpendicular elongated garnets, in our opinion, to permit belief in much movement in most of the rocks since the garnets began to form.

On the other hand, some of the microstructures of inclusions in the garnets appear

to suggest notable rotation of the crystals since growth, and it is necessary to reconcile the two conclusions. Again it seems best to begin with the relations most surely explained—namely, the lines of inclusions in twinned garnets. Many of the garnets are twinned, and the indivuduals are distinguished by distribution of inclusions. Six-rayed aggregates are most common (Pl. 7, figs. 3, 4). In some, each ray is separated from the adjoining ones by a V-shaped re-entrant of abundant graphitic inclusions. Each ray, moreover, though free from the graphitic inclusions, has many tiny parallel lines in the direction of the axis of the ray (perpendicular to the face of the crystal), in which there is a very fine-grained anisotropic mineral, probably quartz. The distribution of graphitic inclusions and the lines of anisotropic mineral are undoubtedly related to crystallographic directions in these garnets. Moreover, in the individuals of the twinned groups, the tiny lines of inclusions are all perpendicular to the growing face.

A second type of inclusion also consists of lines of quartz grains, definitely recognizable but very small. In most sections the lines have different directions in different crystals, but in some sections they are approximately parallel in most crystals (Pl. 7, fig. 5). In many sections in which there is approximate parallelism, the direction is nearly parallel to the schistosity of the rock; but there are nearly as many sections with the lines at high angles, even perpendicular, to the schistosity. In most elongated grains, the lines of inclusions parallel the long axis of the grains; but in some sections the lines are approximately parallel in all garnet grains although the crystals are equidimensional. These lines of quartz inclusions cannot mark an earlier direction of schistosity because the grains are many times smaller than the quartz grains in the rocks of even the outermost metamorphic zone, the biotite zone, and they differ only in size from the lines of inclusions in the twinned garnets, which obviously cannot be inherited schistosity. Both types of lines of inclusions are probably related to crystallographic directions. External forces of compression or shear may, however, indirectly or even directly control the directions of the lines. If the straight lines of inclusions which are nearly parallel in all garnets in a single section are not inherited schistosity, and we think they are not, then it must be concluded that some external force can cause the lines to be oriented during the growth of the garnets at least throughout an area covered by a thin section.

In a few of the sections of the Lead rocks there is a slight curving of the lines of quartz inclusions into faint or even well-marked S-form (Pl. 7, fig. 6). The angular deflection in most crystals is only a few degrees, but in one it was nearly 90°. Harker (1932, p. 220–221) has interpreted similar occurrences, and Dodge (1935, p. 57–58, Pl. VIIa) has interpreted these occurrences, as proof of rotation of the schistosity planes from those in the schists to those in the lines of inclusions. We believe the curving of the lines is proof of rotation of the garnets during growth, but only by the amount of the curving of the lines within the garnets. Straight lines of inclusions are proof of no rotation during growth. Furthermore, they do not prove rotation since formation merely because they make an angle with the schistosity of the rock. Another possible effect of rotation during or since garnet growth is shown by an offsetting of the schistosity planes; but this effect, like the curved lines of inclusions, is very rare in the Lead rocks. We conclude, therefore, that the inclusions in the garnets of

the Lead rocks indicate a little rotation during the growth of some garnets but none during the growth of most of them, and there is no means of measuring the amount of rotation since formation of the garnets. However, absence of positive evidence of rotation of clastic grains of quartz shows that similar absence of positive evidence of rotation of garnets cannot be taken as proof that no rotation has occurred.

Quartz has replaced garnet in many specimens of the Lead rocks, in all stages from a few scattered grains to complete replacement. Moreover, there are all gradations from lines of tiny grains to lines of larger ragged grains and still larger linear areas of quartz in mosaic pattern. Possibly the lines of tiny grains are themselves replacements of garnet by quartz along the tiny lines found usually only in the twinned garnets. On the other hand, some garnets, especially in the staurolite zone and the inner part of the garnet zone, have replaced a small amount of the encircling biotite and quartz in what must be a slight rejuvenation of the period of garnet growth.

Other porphyroblastic minerals give little additional information on the course of the deformation of the Lead rocks. Rotation of staurolite crystals can be detected in some hand specimens by the asymmetric distribution of quartz areas adjacent to the staurolite and elongated in the plane of schistosity. The rotation is, however, not common. Ragged plates of prochlorite are random in orientation in most sections and replace the micas and quartz of the rock. A few crystals elongated transverse to bedding have deformed the micas around the ends of the crystals, just as garnets elongated transverse to bedding have deformed the micas. A very few sections show elongated prochlorite crystals oblique to schistosity and about parallel throughout the section, and these prochlorite crystals have replaced but not deformed the micas of the groundmass. It must be concluded, as was concluded from a study of the garnets, that external forces can cause orientation of the prochlorite crystals throughout the area of a thin section.

The study of the porphyroblasts of the Lead rocks leads to the general conclusion, therefore, that a small number of specimens show rotation of porphyroblasts during growth but most do not, and a small number of specimens possibly show rotation since growth, but for most specimens there has not been much rotation since the formation of the porphyroblasts. It has already been pointed out, however, that some of the cross folding must be later than the formation of the garnets.

This general conclusion, that there has been relatively little movement in the rocks since the formation of the porphyroblasts, except possibly that due to cross folding, is in accord with the study of the zonal pattern of metamorphism in the district (Noble and Harder, 1948), where the boundaries between the metamorphic zones have been shown to be independent of the shapes of the folds.

DILATATION DURING CROSS FOLDING

A structural feature resulting from cross folding may be called dilatation and is believed by us to be a new application of a hitherto little-used concept in structural geology. By dilatation is meant, in this connection, the expansion or inflation of the rock in a zone of cross folding resulting from a reorientation of the minerals. Plate 2 shows that the crenulations in a schist are discontinuous faults of microscopic size and that the well-oriented minerals lose their orientation in the crenulated zones. Since no new minerals are formed, but merely the old minerals are thrown out of

orientation by bending, fracturing, and rotation in the crenulated zones, the conclusion seems inescapable that, as a result of the development of crenulations, the rock has increased in volume.

To answer the obvious question as to the amount of dilatation involved in this mechanism, it is necessary to consider structures larger than these crenulations. Cross folds can be found of every dimension from those of Plate 2 (measured in fractions of an inch) to those of Figure 3 (measured in hundreds of feet). Many folds of sizes visible within the limits of the walls of a crosscut or stope show measurable thickening of stratigraphic units within the area of the cross folding, and on still larger scale the cross folds shown on Figures 1, 2, and 3 show obvious and extensive dilatation of the Homestake formation over and above that due to repetition by folding.

Although the changes in the shapes of folds resulting from dilatation are not important on a regional map, the principle of dilatancy (Mead, 1925) cannot be ignored in comparisons between deformations in schistose rocks and those in more massive rocks which normally fail by faulting on smooth fractures. Moreover, the changes in shapes become of considerable importance in folds of the size of those mapped in mine workings (Figs. 1–3). In a separate paper the effect of dilatation in zones of cross folding will be shown to be of the greatest importance when the question of localization of ore is considered.

RELATION BETWEEN STRUCTURE OF PRE-CAMBRIAN ROCKS AND THAT OF YOUNGER ROCKS

STRUCTURE OF BLACK HILLS UPLIFT

Although the primary purpose of this paper is to discuss the structure of the pre-Cambrian rocks of the Lead district, obviously the relations between the structure of the pre-Cambrian rocks and that of the younger rocks must be considered. The larger features of the structure of the Black Hills uplift have been described by Darton (1901; 1904; 1909), but they warrant repetition because more recent emphasis on the restricted area of the central Black Hills (Darton and Paige, 1925) has tended to overshadow the relations of the larger features. The following generalization of the regional structure has been prepared by the senior author from data in the publications just cited and from other sources. The Black Hills uplift comprises two uplifted blocks (Pl. 4). One, about 100 miles long and 50 miles wide, extends from a point near Edgemont, South Dakota, nearly north to a point a few miles north of Sturgis, South Dakota. The other block, about 100 miles long and 70 miles wide, extends northwest from the northwest corner of the first one across the northeast corner of Wyoming to a point near the southeast corner of Montana. In both blocks the strata are relatively flat on the top but are steep-sided, in places even vertical, near the edges. The strata on the western and southwestern edges are for the most part steeper than those on the opposite edges. Neither "dome" nor "anticline" conveys a wholly adequate description of the shape of these flat-topped, steep-sided blocks. Both blocks are tilted upward gently to the point of intersection in the northern Black Hills. The northward-trending block stands a little higher than the other one and is separated from it by a scarp which strikes a little east of north and passes just east of Newcastle. Either the northward-trending block is the younger, or its uplift has been renewed along the western edge. This edge of the northward-trending

block has been traced southward as a sharp, asymmetric anticline through the Old Woman Creek area and the Hartville uplift, in eastern Wyoming.

Speculation regarding the nature of the forces which caused the uplifting of these blocks must be based on wider studies than can be attempted in this paper. Paige has stated that

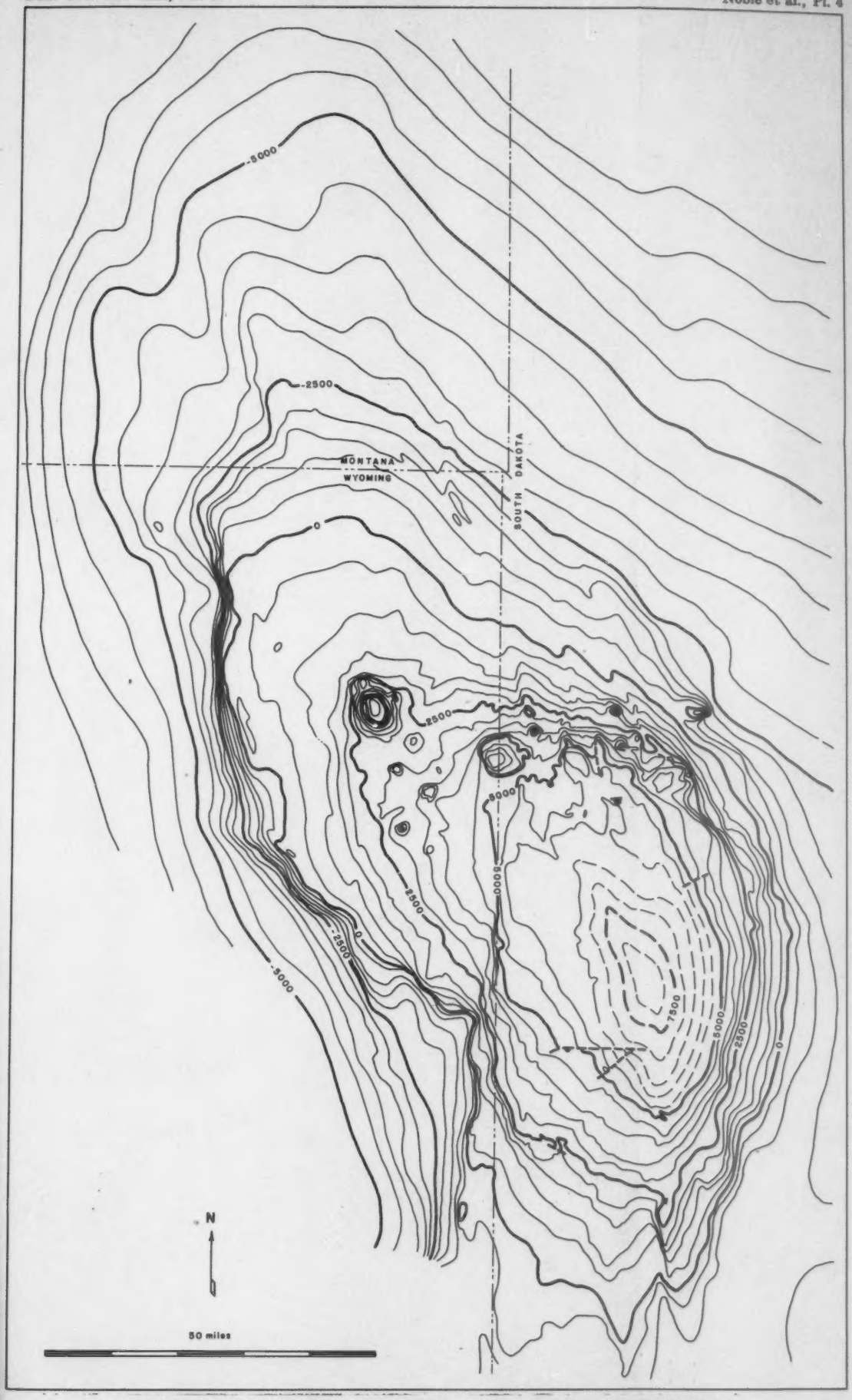
> "these intrusions [i.e., of the Tertiary intrusive rocks of the Black Hills] took place while there was great igneous activity elsewhere. Many great batholiths came into place about this time and produced domes in the beds which they invaded. It is believed that an extensive batholith underlies the Black Hills . . ." (Darton and Paige, 1925, p. 17).

We think it is altogether reasonable to ascribe the uplifting of these blocks which form the Black Hills to Tertiary batholithic intrusion, but we see no possibility of obtaining convincing evidence of the presence of the batholith.

A few minor irregularities mar the symmetry of these uplifted blocks. Near Hot Springs, in the southeastern part of the Black Hills, three small folds diverge from the edge of the main block. Each is flat-topped and steep-sided, steeper on the west side, like the main block. In the northern Black Hills, there are many minor irregularities, all directly related to masses of Tertiary intrusive rock. Circular or slightly elongated domes have cores of Tertiary porphyry. Five small flat-topped folds diverge from the main uplifted block, and each one extends northward from a dome in which is exposed a Tertiary stock or laccolith, but the folds do not extend southward from the associated intrusive bodies. These flat-topped folds, so obviously caused by intrusion, are so similar in form to those near Hot Springs that the question arises whether the Hot Springs folds may be caused by intrusion. No Tertiary intrusive rocks are exposed in the Hot Springs region, however, and there is no indication of the presence of igneous activity except the folds and possibly the hot springs.

The circular or slightly elongated domes and accompanying intrusions of the northern Black Hills have been described by many authors (Russell, 1896; Irving, 1899; Jaggar, 1901; Darton, 1904; 1909; Darton and O'Harra, 1905; Paige, 1913b; Darton and Paige, 1925), all of whom (excepting Russell) classified them as laccoliths, in the sense of the original definition and examples given by Gilbert (1877, p. 18–21, 97, 98). Although Gilbert's definition included any igneous body of approximately lenticular shape injected into sedimentary strata of any attitude, which, by the force of injection, produced an uplift of the overlying sedimentary strata, the typical examples of the Henry Mountains, occurring in nearly flat sedimentary strata, were described by him as having flat lower surfaces and dome-shaped upper surfaces. By most writers the name "laccolith" has been reserved for bodies similar in shape to the typical occurrences described by Gilbert in the Henry Mountains. By this usage a laccolith rests on nearly flat-lying strata or on a nearly flat unconformity. The name has been used with this interpretation by most of the authors just referred to. Paige, however, used the broader definition, with no restriction as to the shape of the lower surface. In recent studies of the Henry Mountains type laccoliths, Hunt (1942) ascribes the main doming to stock intrusion with only subordinate amounts of doming by intrusion parallel to bedding planes.

Some of the Tertiary intrusive bodies of the northern Black Hills may be laccoliths, but some cannot be, in the accepted sense of the definition, and we doubt if many of



CONFIGURATION OF TOP OF PRE-CAMBRIAN ROCKS IN THE BLACK HILLS UPLIFT.

The contours are constructed from folios and other maps by the U. S. Geological Survey, combined with isopach maps by Noble and modified near Lead to agree with mapping by the authors. Contour interval 500 feet. Datum is sea level.

them are true laccoliths. One type of intrusive body is overlain by a dome of pre-Cambrian and Paleozoic rocks. Examples are the Bear Lodge or Warren Peaks uplift in Wyoming, the Nigger Hill or Tinton uplift near the boundary between Wyoming and South Dakota, the schist area of the Lead district and the adjacent

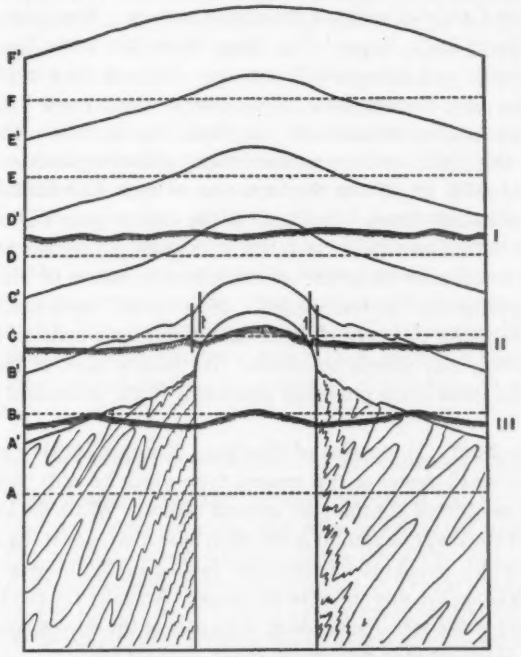


FIGURE 4.—*Diagram of pseudo-laccolithic doming of flat-lying strata*

Rocks adjacent to intrusive body have been separated by the force of intrusion. In addition to sharp uplift localized over the intrusive body, there is a broader doming caused by the added mass of intrusive rock. Lines A', B', C', D', E', and F', represent stratification planes originally at A, B, C, D, E, and F. In the rocks near the walls of the intrusive mass there is compression, and there is probably shear in addition, due to viscous drag by the rising magma. In the rocks over the top of the intrusive mass there is tension. Dikes and small intrusive masses will probably fill any tension cracks that may form. Hypothetical erosion surfaces are shown at I, II, and III. At stage I a dome, possibly cut by dikes, might be interpreted as covering a hidden laccolith. At stage II a circular mass of intrusive rock is surrounded by outward-dipping sedimentary rocks, somewhat faulted. It would not be possible to distinguish the stock from a laccolith at this stage. At stage III a circular mass of intrusive rock is surrounded by a border of schist, which in turn is surrounded by outward-dipping sedimentary rocks. This stage does not permit classification of the intrusive body as a laccolith. If the intrusive stops short of the horizontal strata, those strata will be domed, and the schists between them and the intrusive mass will be uplifted and highly deformed.

districts, and the small uplift of Whitewood Peak, a few miles northeast of Lead. The pre-Cambrian rocks are sedimentary rocks and foliated granites and amphibolites, all with steeply dipping foliation planes. It seems impossible, therefore, to conceive that the intrusive bodies have any close approximation to the ideal laccolithic shape of the original definition by Gilbert. In all probability they are stocks with nearly vertical walls, whose bulk has caused a doming of the adjacent and overlying rocks (Fig. 4). The laccolithic mechanism offers no explanation for a notable

uplift of the pre-Cambrian rocks except that of a floating of large bodies of pre-Cambrian rocks in the porphyry magma (Darton, 1904, p. 8; Smith, *in* Darton and O'Harra, 1905, p. 4; Ferguson and Turgeon, 1908), a mechanism which is not only highly improbable but is refuted by the field relations.

A second type of dome is surrounded by steeply folded and partly faulted Paleozoic strata but no pre-Cambrian rocks at the erosion surface. Examples are Bear Butte, Crow Peak, Citadel Rock, Ragged Top, Sheep Mountain in the Bear Lodge Mountains, Black Buttes, and Inyankara Mountain. Some of these uplifts are bounded by faults on one, two, or even three sides. Paige (Darton and Paige, 1925, p. 23; Fig. 31) believes that the intrusion of Crow Peak "forced its way through the overlying strata at this place, much as a punch might perforate plastic material", a process which would seem to call for the formation of a circular fault in the overlying Paleozoic and Mesozoic strata. In Figure 4, the erosion stage II is a close approximation to that of the Crow Peak "laccolith". Russell (1896) had previously postulated a similar mechanism for several of the intrusive masses of the northern Hills, which he characterized as "plutonic plugs". Here again it seems doubtful if the name laccolith is suitable for this type of intrusion although the uplifts with only a small amount of faulting may overlie laccoliths. We believe most of the uplifts of this group differ from those of the preceding group only in the position of the erosion surface.

In the Little Rocky Mountains of Montana, Knechtel (1944) has mapped and described many small domes of this general type which he calls "trap-door blocks" and attributes to upward punching by igneous bodies. While the analogy to a trap door is good for the shape in plan, it is not good in section; nor is the mechanism that of a trap door, which would call for concentric folding on the hinge side and an opening of the rocks along the side opposite the hinge. Actually the mechanism is that of upward shearing of the subcircular block, with part of the movement along fault surfaces and part along zones of distributed shear, or shear folds.

In a third group of domes, Paleozoic and Mesozoic strata dip outward at low angles from central masses of intrusive porphyry. Faulting has not been found in most of these and probably is not common. Examples are Vanocker Mountain and Sheep Mountain (Mt. Theodore Roosevelt). There are also a few smooth domes of Paleozoic and Mesozoic strata which are probably underlain by intrusive masses not yet exposed by erosion. Examples are Elkhorn Mountain near Whitewood and several small domes in the Sundance quadrangle. There seems to be no more reason to ascribe these domes to laccoliths than to stocks.

A fourth group of intrusive masses and intruded Paleozoic rocks, called laccoliths by Paige (Darton and Paige, 1925, p. 19–23), probably are for the most part composite sills and dikes separated by small partings of sedimentary rocks, principally shale of the Deadwood formation. A few may be true laccoliths, but the feeding conduits are probably dikes rather than vertical pipes. The examples, too numerous to tabulate here, are distributed throughout the triangular area bounded by Custer Peak, Sturgis, and Spearfish.

STRUCTURE OF UPLIFT BETWEEN DEADWOOD AND TROJAN

One of the domes of the northern Black Hills includes the Lead district (Pl. 1). The dome extends from Deadwood to Trojan, near the northwest corner of the north-trending uplifted block, and is one of the points of greatest uplift. The dimensions of this dome are about 12 miles by 10 miles, the longer dimension extending about west; but a flat-topped fold which extends to the northwest for several miles may be a part of the same uplift. Near the center of the dome stands a very ragged stock, elliptical in outline and about 2 miles wide and 4 miles long. This is the Cutting stock (Figs. 5, 6). There are several other stocks, smaller and more ragged in outline, and vast numbers of dikes, large and small. Pre-Cambrian rocks surround the Cutting stock, but they are not quite symmetrical around it. They extend farther east than in other directions, partly because of deeper erosion by Whitewood Creek and its tributaries in that region, but partly because of a second uplift on the Sheep Mountain stock near the eastern edge of the area. The Lead district (Pl. 1) includes most of this extension of the pre-Cambrian rocks to the east. Surrounding the pre-Cambrian rocks is the outward-dipping Cambrian Deadwood formation, and in the Deadwood are many sills of Tertiary porphyry. The younger Paleozoic rocks, beginning with the Ordovician Whitewood formation, surround the outcrop of the Deadwood formation, except that to the southeast of the dome they have been removed by erosion.

EFFECT OF TERTIARY DEFORMATIONS ON PRE-CAMBRIAN ROCKS

It can scarcely be doubted that the Tertiary intrusions locally caused notable deformations in the pre-Cambrian rocks and may have caused widespread minor deformations. The reconnaissance mapping of the dome between Deadwood and Trojan shows a close relationship between intrusions of Tertiary porphyry and the doming and faulting of the Paleozoic rocks. On the southern edge of the pre-Cambrian area, the Cambrian Deadwood formation is domed and faulted so that its outcrop swings in a semicircle around the Cutting stock (Figs. 5, 6). The faults strike a little west of north, and each displacement augments the doming. Dikes of Tertiary porphyry have been intruded along many faults, and there are many more dikes parallel to the faults but not occupying the faults. (The smaller dikes have been omitted from Figures 5 and 6.) Similarly on the northern edge of the schist area doming is accompanied by faulting. Near the western side of the pre-Cambrian area there is more doming and faulting, some of which is obviously attributable to the intrusion of Tertiary porphyry. All this doming and faulting has been measured on the Cambrian strata. Obviously the pre-Cambrian rocks have been deformed at least as much as the Paleozoic rocks, and possibly more. The pre-Cambrian rocks have acted as a cushion between the rising masses of magma and the overlying Paleozoic and Mesozoic strata.

Although these faults which are so plain in the Paleozoic strata are recognizable in the pre-Cambrian rocks in only a few places and in those places with no great certainty, other deformations of the pre-Cambrian rocks can be certainly attributed to intrusions of Tertiary porphyry. The most conspicuous deformation is that which produces, in the northeastern part of the Lead district, north and northeast strikes in

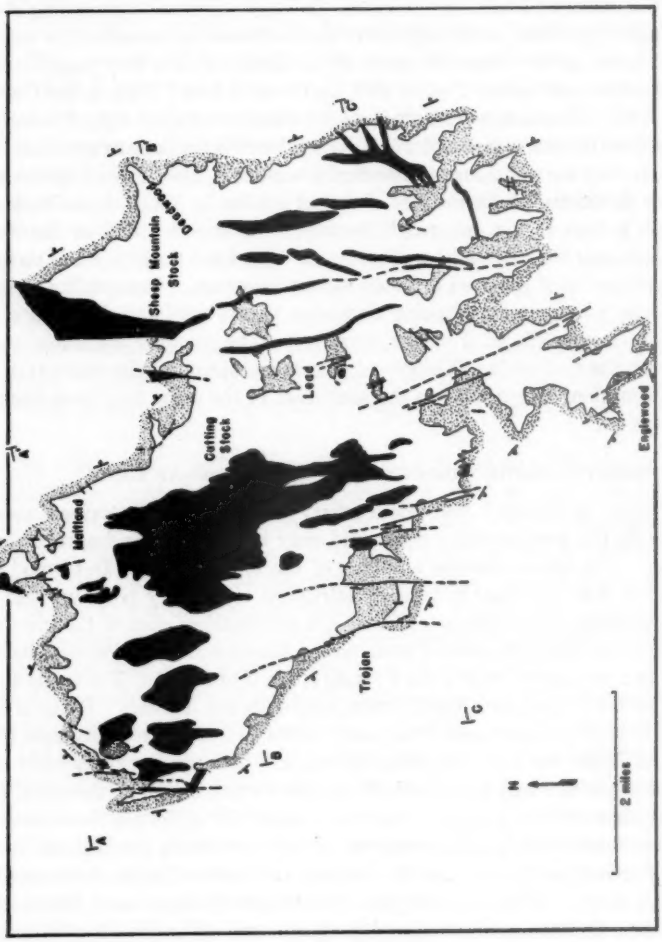


FIGURE 5.—*Map of the uplift between Deadwood and Trojan*

Sills in Cambrian strata and small dikes in pre-Cambrian rocks omitted. Most faults in pre-Cambrian rocks omitted. Cambrian strata stippled. Tertiary intrusives solid black. Faults shown by dashed lines. Pre-Cambrian rocks occur inside the edge of the Cambrian strata.

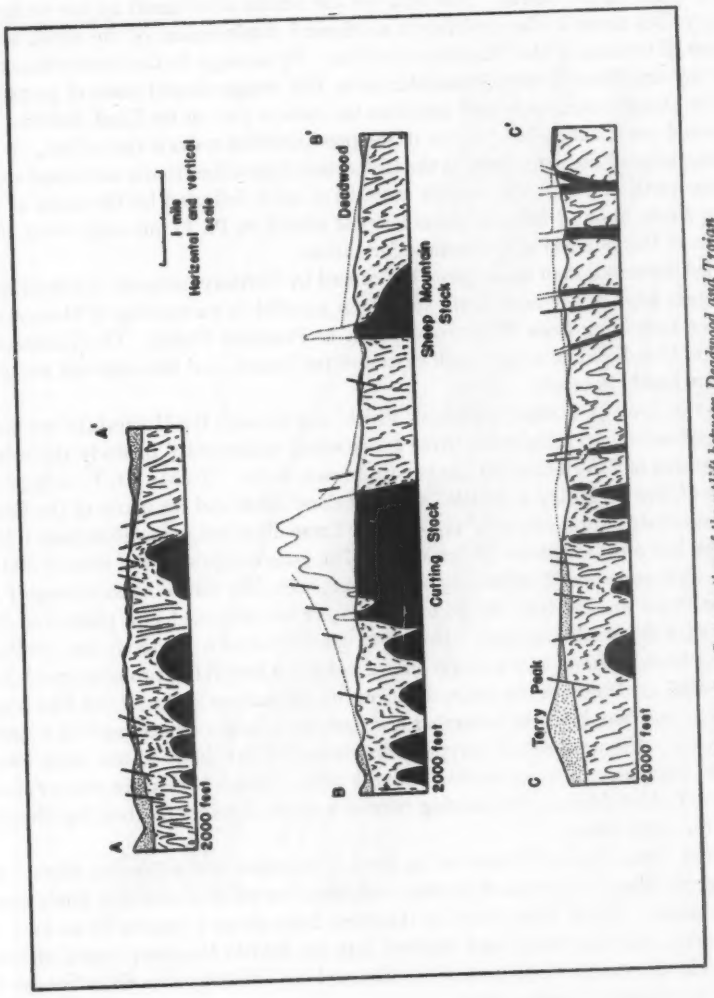


FIGURE 6.—*Cross section of the uplift between Deadwood and Trojan*
Conventions are the same as in Figure 5.

place of the usual northwest strikes (Pl. 1). It seems most reasonable to suppose that the pre-Cambrian structure is deflected around the large wedge-shaped mass of porphyry of the Sheep Mountain stock, which is just northeast of Blacktail Gulch and is much larger farther north. Not only are the schists pried apart by the wedge of porphyry, but there is also probably a southward displacement of the whole schist area east of the axis of the Caledonia syncline. By analogy to the known structure of the rhyolite dikes in the Homestake mine, this wedge-shaped mass of porphyry probably plunges southward and underlies the eastern part of the Lead district.

A second conspicuous deformation of the pre-Cambrian rocks is that which, on the northwest edge of the map, deflects the Homestake formation from a northwest strike to about north (Pl. 1). The schists must have been deflected by the mass of the Cutting stock, about 2 miles in diameter (not shown on Pl. 1) but only about 1000 feet west of this outcrop of Homestake formation.

A third deformation of schist probably caused by Tertiary intrusion is a fault near the western edge of the map along and about parallel to an outcrop of Homestake formation extending from Whitewood Creek to Poorman Gulch. The Homestake formation is reduced to a few small disconnected lenses, and the adjacent rocks in places are highly sheared.

A zone of dikes of Tertiary porphyry which cuts through the Homestake ore body and is exposed on all mine levels gives an excellent opportunity to study the small-scale features of deformation of the pre-Cambrian rocks. To Mr. D. T. Griggs belongs credit for beginning a detailed study of these dikes and for many of the interpretations of structural effects of the dikes. These dikes range in width from 1 foot to 20 feet, but most are about 10 feet wide. The zone comprising the dikes is 300 to 400 feet wide and about 3 miles long. In shape, each dike can be likened roughly to a human hand so held that the palm and fingers are extended in a plane and the fingers point diagonally upward. If a given dike is cut at a relatively low position by a mine level, it shows only a single dike of unknown length but probably much less than 3 miles, the length of the entire dike zone on the surface. But, if the dike is cut at a higher position, it shows a number of fingers with long axes arranged in a single straight line. Figure 7 shows a typical occurrence of the dikes on one mine level. The dikes were injected in succession, and new dikes commonly came in near or along the walls of older dikes, either leaving narrow screens of schist or abutting directly against the older dikes.

So-called "breccia dikes" attest to the force of injection of the rhyolite dikes. In these breccia dikes, fragments of rhyolite and schist are set in a matrix of finely comminuted schist. These dikes range in thickness from about a quarter of an inch to about 6 feet, and they have been injected into the schists for many scores of feet. The breccia dikes carry fragments of rhyolite and cut older rhyolite dikes but are in turn cut by younger rhyolite dikes.

Most rhyolite dikes show no definite fracture patterns. A few wide dikes show columnar jointing in the central portion, the columns standing at right angles to the dike walls. In some parts of some dikes, joints strike across the dikes and dip steeply north, which makes them about normal to the direction of elongation of the dike fingers. Some dikes show some jointing parallel to the walls, in most dikes only in

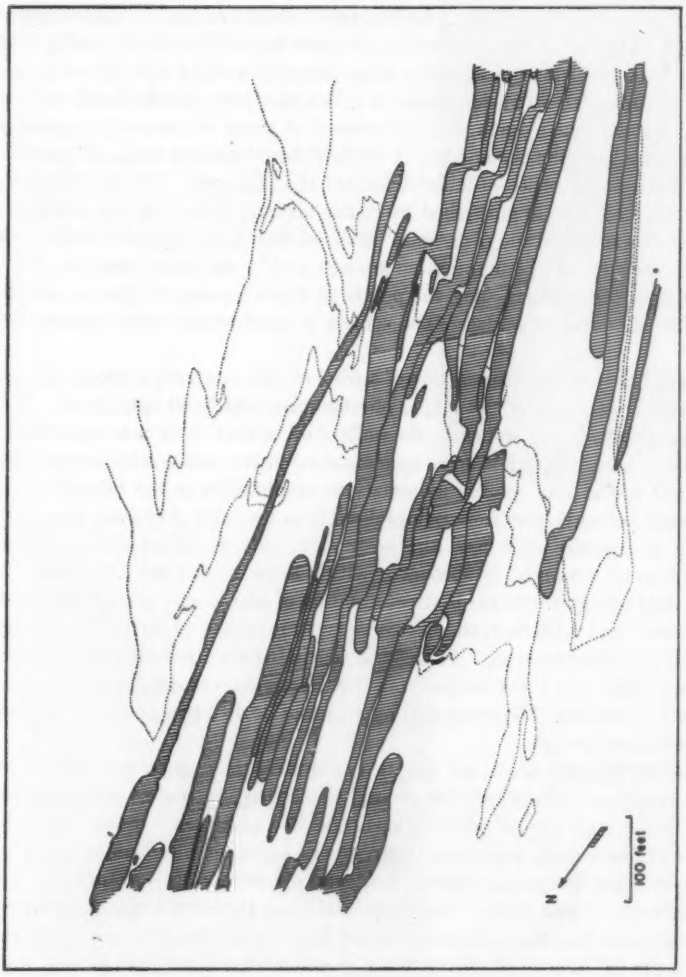


FIGURE 7.—*Map of the dike zone on the 2000 level, Homestake Mine*
Rhyolite dikes cross hatched. Outline of Homestake formation dotted.

the last few inches against the contact. Tiny crystals of pyrite, calcite, and in places fluorite line the fractures.

In a few places the dikes taper to a relatively narrow point, but much more commonly the dikes, irrespective of size, end bluntly, on a nearly semicircular curve. A study of the structure of the schists at the blunt semicircular ends of the dikes suggests that the dikes have been forcibly injected and have shoved the schists bodily aside and upward. The schists in most places wrap smoothly around these blunt ends of the dikes. In some instances a few inches of schist have been attacked and removed in some way by the dike magma, but the amount of schist so removed is probably everywhere much less than the amount of the introduced igneous rock. Figure 2 of Plate 6 is a typical illustration of schist deformed at a dike end. In a few places the bending of the schists is accomplished by minor faulting (Pl. 5, fig. 1), but this is probably not common. Many dikes show offsets in their walls, opposite walls of the dike usually being offset about the same amount and in the same direction (Pl. 5, fig. 2). Many of these offsets mark the points at which separate fingers of the dike merged as the dike was extended. The schists in most places wrap around these offsets also.

Microscopic study of the deformation of schists at dike contacts supports the conclusions reached by study of the maps, photographs, and hand specimens. Most sections of specimens from points near the walls of dikes show little or no deformation of pre-dike structure although in some specimens extensive replacement by post-dike carbonate and pyrite may conceal deformations attributable to the intrusion. On the other hand, all specimens from points directly in line with dike ends show some deformation of pre-dike structure, and some show intense deformation. Certain evidence is limited, however, to specimens within a few feet of the dike ends. Although the field relations indicate that deformation of schists may extend for a much greater distance than a few feet, the microscopic study makes no distinction between pre-dike and post-dike shearing. It must be concluded that beyond a few feet from the dike ends, and even a few inches from the dike walls, the deformational effects attributable to intrusion of Tertiary porphyry cannot be distinguished from deformations of pre-Cambrian age.

Most dike endings and offsets are wrapped around by the schists, but three or four round dike endings or offsets have been found which sharply truncate the schist bedding with little or no visible deformational effects. Considered alone, these dike occurrences can be held to support a theory of emplacement by assimilation of wall rock. The evidence from other sources, however, favors forcible injection with little or no assimilation of wall rocks. Probably, therefore, these few examples were also injected, and schist has been displaced wholly by slipping along schistosity planes.

Although the deformational effects of the injection of a single dike are not great, the sum of the deformation effects of all the dikes of the dike zone may be of some magnitude. Where the dikes are continuous and of uniform width, the deformational effects on the schists were mainly to move them apart by the width of the injected igneous material. The total width of igneous material in the central portion of the dike zone is about 200 feet, measured perpendicular to the dike walls, and the schists have been shoved aside by this distance.

FIGURE 1. DEFORMATION OF SCHIST
Scale is 6 inches long. Schists are bent around dike end and are also faulted slightly. Displacement almost wholly to one side, which is the footwall side of dike. Taken under direction of D. T. Griggs.

FIGURE 2. INTERMEDIATE STAGE IN FORMATION OF DIKE OFFSET
The wider dike finger is about 3 feet wide. The two dike fingers are just barely joined. Schist is bent sharply to wrap around original positions of dike fingers. Taken under direction of D. T. Griggs.

DEFORMATION OF SCHIST BY DIKES

FIGURE 1. EAST WALL OF HOMESTAKE OPEN CUT
Iron-stained dark chlorite schists of pre-Cambrian Poorman formation overlain by thick sill of Tertiary quartz porphyry (light color, with columnar jointing) and several sills of Tertiary rhyolite (light color, with almost no jointing). A little nearly flat conglomerate, quartzite, and dolomite of Cambrian Deadwood formation (brownish gray) rests on schists and intervenes between sills. Dikes of Tertiary rhyolite (light color) cut the schists near right edge of view and turn to make sills in the Deadwood formation. Dikes dip east, away from camera. Base of Deadwood formation is displaced where it crosses each dike.

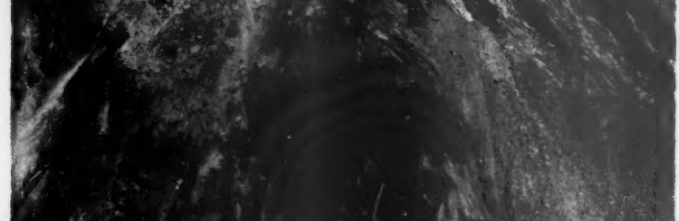


FIGURE 2. MOST OF END OF DIKE FINGER EXPOSED ON WALL OF CROSSCUT
Left edge of dike extends about 2 feet beyond field of camera. Scale is 6 inches long. Schists are in about normal pre-dike attitude on right edge of view, but over dike end they are deflected by upward push of dike magma. Small lenses of quartz conformable to bedding, probably originally chert, are bent but for the most part are not fractured. Tiny seams of late calcite cross folded schist and quartz lenses.

In the exposure of the intersection of the nearly flat base of the Deadwood formation and the dike zone in the Homestake open cut we can measure the displacement of the base of the Deadwood formation in crossing the dike zone (Pl. 6, fig. 1). Mapping inside the open cut shows that the total vertical displacement of the base of the Deadwood formation of 112 feet is made up of displacements across each dike, in amounts exactly attributable to a separation of the walls at right angles to the plane of each dike.

An approximate representation of the pre-dike structure of the schists can be obtained by graphic reconstruction, by assuming that all the igneous material formed a single dike with a width equal to the total width of the dikes and an attitude corresponding to the average attitude of the dikes, and that the movement was normal to the walls. A more accurate representation of the pre-dike structure can be obtained by a graphic method consisting of taking out each dike separately and fitting the schists of the opposite walls together again. Maps and sections reconstructed by this method do not, however, give the true pre-dike structure. The graphic methods assume that the individual dikes are of indefinite extent and uniform width and that the wall rocks are rigid and of uniform strength. It seems probable that eventually the lateral spreading of the schists was transposed into an upward bulging; the broad effect of the injection of the dikes probably was a slight arching at the surface. There seems to be no way to determine how far the lateral spreading extended. A curving of the De Smet synclinal axis where it approaches the dikes on the 2600 level (Fig. 1) may indicate that the lateral spreading ends rather suddenly. Longitudinal sections show some steepening of plunge of structures on approaching the dike zone and a corresponding flattening after crossing the zone; this may also point to a disappearance of the effects of lateral spreading.

Where the dike injection has formed a complicated pattern of dikes and fingers, segments of schist between dikes have been considerably distorted and displaced. In some parts of the dike zone on the 2000 and 2150 levels, where the pattern is complex and mine exposures are excellent, the graphic method of reconstruction of the pre-dike structure was only partly successful. Certain blocks of schist surrounded by dike material clearly have been greatly deformed and also have been carried bodily many scores of feet. In many places dike fingers have been injected into a narrow band of deformed schist separating two earlier dikes, thereby further deforming and in places bodily transporting the intervening schist but not fracturing the earlier dikes. In some of these places the injection of the dike fingers into the deformed schist between the rigid earlier dikes has produced an effect not unlike that of a piston operating on plastic material confined between rigid walls.

Another deformational effect which prevents a precise graphic reconstruction of the pre-dike structure results from the unequal strength of different types of schist. The ore-bearing portion of the Homestake formation, for example, was very extensively chloritized prior to injection of the dikes and was weaker than the underlying Poorman formation or the overlying Ellison formation. The reconstructed maps of some levels show a pronounced reduction in the area of the Homestake formation where that formation crosses the dike zone.

STRUCTURAL HISTORY OF THE LEAD DISTRICT

Although many phases of the structural history of the Lead district remain obscure, certain phases can be correlated with some certainty. Undoubtedly the pre-Cambrian sedimentary rocks were first closely folded into narrow, straight isoclinal folds and later were refolded into new shapes. The intrusions of gabbroic magma which are now represented by amphibolites came late in the period of isoclinal folding. The formation of garnets, a single main episode in spite of a little evidence of rejuvenation of garnet growth, probably intervened between the isoclinal folding and the cross folding. Elongated garnets in random orientation are probably too abundant to have formed before the end of isoclinal folding. On the other hand, in some cross-folded schists the crenulation planes attributable to the cross folding are localized by garnets. It is of course possible that there was more than one period of cross folding. Another reasonably definite age relationship is that the rhyolite dikes of the Homestake mine cut the cross folds. Some faults approximately contemporaneous with the dikes also cut the cross folds. Inasmuch as the Tertiary intrusions were injected not in a single episode but in succession, the age relation between cross folding and all Tertiary intrusion is not determined, especially as the rhyolite dikes formed late in the intrusive period (Darton and Paige, 1925, p. 16).

An undisclosed batholithic intrusion of pre-Cambrian age northeast of the Lead district has been postulated (Noble and Harder, 1948) to explain the metamorphic zoning of the district. Whether this batholithic intrusion had any relation to the period of isoclinal folding is only conjecture. That it can be the cause of the cross folding seems improbable in view of the evidence that cross folding is later than garnet growth, unless there was more than one period of cross folding. The batholith may, however, be the cause of the progressive steepening of the plunges of isoclinal folds from the southwest side to the northeast side of the district.

One of the most perplexing questions of the geology of the Lead district is whether or not the cross folding is the result of Tertiary intrusion, and this question cannot be answered satisfactorily at this time. The cross folding is probably later than the metamorphic zoning but earlier than the rhyolite dikes. No force has been suggested as the cause of the shearing whereby the cross folds were produced, but the relations between stress and strain in rocks must be very direct and obvious to be convincing. The stresses which were transmitted through the schists from rising masses of Tertiary magma and which arched and faulted the overlying strata of Paleozoic and Mesozoic rocks may have been adequate to produce the cross folding of the schists. The major deformations are attributable to shearing over broad zones. The minor features of the major deformations are microscopic shear zones, lines of bending or crenulation of the schist minerals, and in places recrystallization and reorientation of sericite. In Eocene time, when the uplifting of the Black Hills first commenced, the pre-Cambrian rocks were covered by about 8000 feet of strata. To deny that recrystallization of sericite could occur in Eocene time is to deny the existence of adequate pressure, since clearly temperature, time, and solutions may have been adequate. It may be that pressure due to a column of 8000 feet of rock is inadequate to permit recrystallization of sericite; but, under the conditions prevailing in the Lead

rocks during the injection of Tertiary magma, pressure was measured not by the weight of a column of rock but by the probably much greater force of the resistance of the whole column to rupture by shear.

A determination of the time relations of cross folding and Tertiary intrusion cannot prove that the cross folding was not caused by these intrusions. A little study will show that the deformation of wall rocks by forcible injection of a mass of magma must appear to be earlier than the actual intrusion. When a result travels in front of a moving cause, the result necessarily antedates the cause at a given point. The Tertiary intrusions were injected into rocks already folded and faulted by the force of the injection, and the deformation, though caused by the igneous intrusion, preceded the intrusion.

It is not even possible to be sure whether the cross folds have deformed the Cambrian strata. At first glance, it seems obvious that the amount of Tertiary deformation of the Cambrian strata is much less than the amount of displacement in the large cross folds of the mine area, but the relations may not be so simple as that. The large cross folds are probably discontinuous, like the small folds and crenulations; so the displacement on a given cross fold probably is at a maximum in the central part of the fold but decreases to zero at each end. Moreover, crenulations and small cross folds are developed only in the schistose rocks, and application of the same stresses to the stronger Paleozoic rocks probably resulted in broad bending and minor faulting. The base of the Cambrian strata where it crosses the limits of known structures in the mine area has been tilted upward by uplift on the west side, and in addition it has been bent into small terrace-shaped folds also with the west side up. This tilting and bending correspond in direction and may correspond in amount to the displacements on the cross folds in the pre-Cambrian rocks.

Uplift of the Black Hills by several thousand feet since Oligocene time is inferred (Darton, 1901, p. 559–561; Fenneman, 1931, p. 85) by the presence of Oligocene gravel deposits in the Lead district, nearly on the highest part of the Black Hills uplift. These outcrops are about 3000 feet above deposits of the same age in the White River Badlands, 60 miles east of the Black Hills. These facts actually prove, however, that very little change in elevation of the Black Hills can have occurred. The gravels of the Lead district are alluvial deposits of local origin. The Oligocene deposits of the White River Badlands are in part alluvial and contain some boulders of schist and quartzite of considerable size very probably transported from the pre-Cambrian areas of the Black Hills. The overall gradient from the pre-Cambrian areas of the Black Hills to the Oligocene deposits of the White River Badlands is about 50 feet per mile, which is within the range of gradients of rock-pediment surfaces and alluvial fill in some Arizona streams to-day (Gilluly, 1937, p. 338). The slight amount of rejuvenation shown by some of the main streams and tributaries of the Black Hills may be attributable to a moderate amount of uplift in recent time, to a lowering of base level, by a number of causes, or to climatic changes. In any case, very little uplift would be demanded. We believe the evidence indicates that there has been little or no vertical movement of the area comprising the Black Hills since Oligocene time.

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Calif. Inst. Technology, Pasadena, Calif.; Homestake Mine, Lead, S. D.; Homestake Mine, Lead, S. D.

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PLATE 7.—PHOTOMICROGRAPHS OF PORPHYROBLASTS
× 50

Figure

1. Deformation of bedding by a garnet. A twinned garnet spreads the bedding planes in a biotite-garnet schist. The deformation is interpreted as being due to the growth of the garnet crystal. Plane polarized light.
2. Deformation of schistosity by growth of a garnet crystal. An elongated garnet has grown at about right angles to the schistosity and by its growth has deformed the schistosity. The manner in which the micas wrap around the lower end of the garnet indicates that the deformation is not due to rotation by slipping on the schistosity planes. Plane polarized light.
3. Inclusions in twinned garnets. A six-sided twinned garnet is set in a matrix of fine grained quartz and graphite. Inclusions of graphite separate the individuals of the twinned crystal, and lines of small quartz inclusions are perpendicular to each of the 6 crystal faces. Plane polarized light.
4. Same field as 3, but with crossed nicols.
5. Parallel lines of inclusions in garnet crystals. The inclusions of quartz in the garnets seem to be exactly parallel in these grains, but the parallelism is not exact over the whole thin section. Plane polarized light.
6. Rotated garnets. Lines of small quartz inclusions in garnet crystals are in S-form, due to rotation of garnets during growth. Only two or three examples of this type of rotation have been found in the Lead specimens. Plane polarized light.

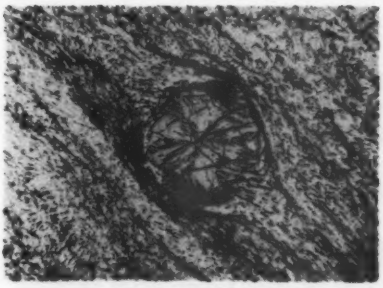

FIGURE 1

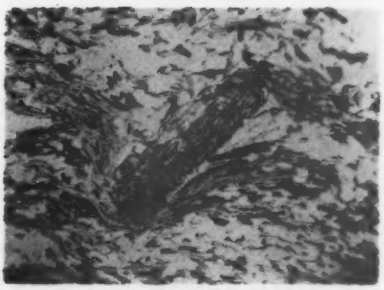

FIGURE 2

FIGURE 3

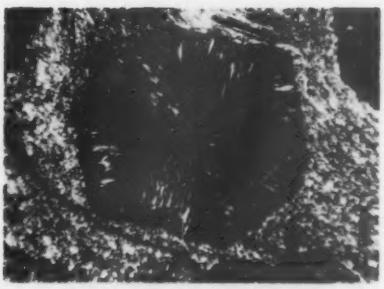

FIGURE 4

FIGURE 5

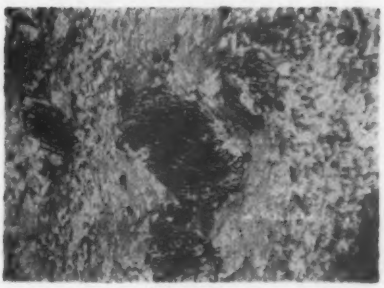

FIGURE 6

PHOTOMICROGRAPHS OF PORPHYROBLASTS

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VOL. 60, PP. 353-362, 3 FIGS. FEBRUARY 1949

ORIGIN OF THE ASYMMETRICAL SHAPE OF THE EARTH'S SURFACE AND ITS CONSEQUENCES UPON VOLCANISM ON EARTH AND MOON

BY B. G. ESCHER

(Presidential address, International Association of Vulcanology, 1948)

When we study the big problems of regional geology, the distribution of mountain chains or tectogenic belts, the distribution of epicenters of shallow or deep earthquakes or seismic belts, or the distribution of volcanic belts, we always debouch into the still greater problem of the continents and oceans. As geologists we are aware that the first named problems can only be tackled with any chance of success if we start from the exact solution of the biggest of geological problems: the remarkable division of the earth's surface in continents and oceans.

As big as this problem is, its solution is as difficult. Geologists and geophysicists in all countries are working on it. Nobody can comprehend all the facts which must be taken into consideration for a satisfactory solution. Therefore every geologist or geophysicist departs from a different combination of data, and so the answers must become dissimilar.

Naturally with such a problem it is necessary to schematize the phenomena. It is not astonishing that geophysicists who enter upon this problem are compelled to perform this schematization to a high degree, as otherwise the problem would not be suitable for a mathematical treatment.

There are however few geological problems which are suitable for mathematical treatment. Most of them are too complex and contain too many unknown quantities. In order to make mathematical treatment possible, a choice must be made of unknowns which are to be eliminated. Now the danger arises that this choice is not made by geological but by physical insight. The result in these cases will be of little geological value.

Of course the geological solutions are likewise often of poor value; but they announce themselves as ordinary actions of thinking, without the gilt frame of mathematics.

Still the opinion is current that a geological paper with mathematics is of more value than one without.

Please do not think that it is my purpose to declare that geophsyics has not offered great service to geology. In the realms of seismology and gravitation, for instance, geology owes a great deal to geophysics. But as soon as geophysics loses touch with geology, the danger arises that its results will be barren.

The problem of continents and oceans contains three main morphological facts, namely (1) the appearance of two levels on earth, the continental 875 m above and the oceanic, 4120 m below the sea level; (2) the ratio of the areas which G. K. Gilbert already in 1892 called the continental and oceanic plateau, is 1:2 (Littlehales, 1932,

mentions the following data: Oceans minus shelves 335 × 10^6 sq. km. or 66%, continents plus shelves 175 × 10^6 sq. km. or 34%) and (3) the arrangements of the continents on the earth is asymmetrical. Moreover we have to deal with the geological problem of the sialic composition of the continents which float in a simatic

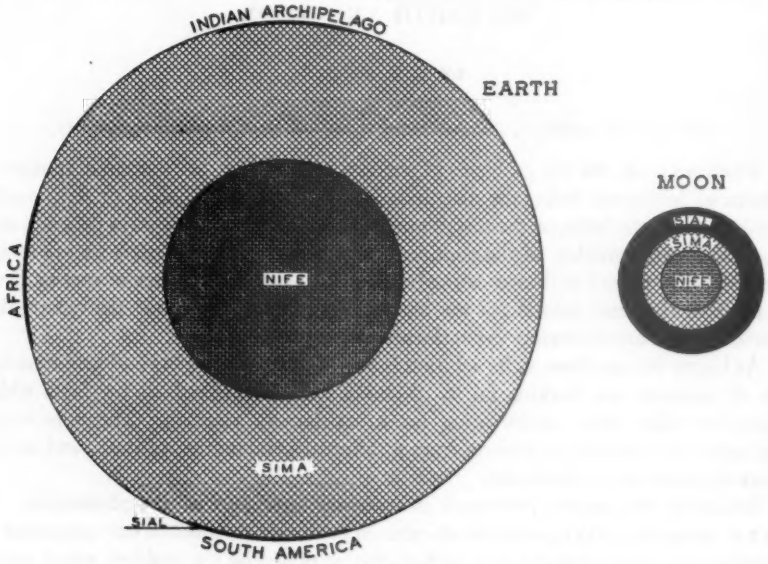

FIGURE 1.—*Hypothetical cross sections through the earth and the moon*

substratum. The theory of isostasy in combination with petrology has elucidated this conception.

In latter years several attempts were made to explain the distribution of continents and oceans by means of forces operating within the earth. Subcrustal currents, generally convection currents, were thought to be the motor of this distribution. It is possible to account in this way for the origin of the two main levels on earth, our first fact.

In 1944 (also in 1948) Vening Meinesz gave a theory in this direction. But his results are not in agreement with our facts 2 nor with 3. For his two levels have the same area and their distribution is symmetrical. The newest contribution in this field from G. F. S. Hills (1947) is able to explain our problems 1 and 2 by coupling the origin of the high level to the birth of the sial, a consideration which seems to me important, especially from a petrogenic point of view. But he does not consider, let alone, elucidate, our problem 3.

In a floating and revolving body such as the earth after its birth, we will have to start from a liquid sphere which was heterogeneous in radial direction, the density increasing inwards, but homogeneous in tangential direction. Hills lets the sial come into existence by means of convection currents. From the two possibilities:

"small currents over the whole surface of the earth" and "an organized convection system," he chooses the second. The first would produce "a uniform cover of granite over the whole surface of the earth." But he sees no way to bring this supposition into accordance with the "actual distribution of granite." I fully agree with him that this distribution cannot possibly be explained by forces acting within the earth. I pointed to this conclusion in 1946 (p. 5, 6, 20) when I declared the idea of Umbgrove (1942, "A new hypothesis", p. 104–110) about this subject unacceptable. Substantially, Wegener (1929, p. 208–209) had already expressed this idea. In the second edition of his book (1947) Umbgrove has rewritten the chapter regarding this matter, but he sticks to the notion that the continents were formed by wrinkling of the primitive thin sialic skin, which covered the whole earth.

Hills gives an explanation for the formation of two sialic continents at the poles, which he calls Laurasia and Gondwana; but he does not explain how these two vast continents changed into asymmetric distribution of continents which now exists.

If we assume that these two polar continents were torn to pieces by convection currents driven by the heat from the radioactive elements in the sial, it is not to be expected that such an irregular configuration would originate as we now are acquainted with. Holmes (1929; 1945, p. 508–509) has interpreted the actual shape and distribution of the continents in this way. By so doing we must suppose that the horizontal forces exercised by currents in the sima upon the sial were so great as to tear the two sial continents Laurasia and Gondwana to pieces and to transport the blocks through the sima which was already solidified in its outer parts.

It is especially our third problem, that of the asymmetrical face of the earth, that forces me to seek the solution in another direction, that in which, already in 1882 Fisher envisaged our problem. He republished this view in his "Physics of the earth's crust" (1889, p. 336–341). It appears to me that only a catastrophic influence from without the earth can have given it its asymmetrical face. This catastrophe was the separation of the moon from the earth.

If we assume this starting point for the explanation of the asymmetrical face of the earth the following are implied:

(1) The earth must have gone through a differentiation into sial and sima before the separation of the moon. In order to explain this point, we should suppose not the second but the first alternative of Hill: "small currents over the whole surface of the earth." By this process the earth would get an exterior shell of sial.

(2) We must assume that the outmost part of the earth was solidified before the separation of the moon took place. Otherwise a scar would not have remained on the face of the earth after the great event.

If one does not accept these two consequences, the following consideration is of little value. I accept them because different phenomena on the earth as well as on the moon find them a logical explanation (Escher, 1939, 1940, 1946).

According to Jeffreys (1929, p. 79) the moon, if it was ever part of the earth, was separated from it within 10,000 years of the formation of the earth, "about the time when solidification" of the earth "was starting." So it must have been in the very beginning of the about 3000 millions years (Holmes, 1946, 1947) long history of the earth. About two thirds of the sialic skin was peeled off the earth, while also a

part of the underlying sima was taken away, as well as a part of the material which forms now the core of the earth. We shall leave undecided what is the nature of the material now forming the earth's core. In our problem this is not of immediate interest.

At that moment only the uppermost layer of the sial was frozen. Mixed up with great quantities of fluid magma, it was remelted in the sphere that was to form the moon. Or did perhaps some floes escape the process of melting down and now form those never understood parts of the moon which are called mountain ranges?

A computation shows that if two thirds of the sial was removed to the moon, our satellites must possess a sialic skin 11 times as thick as it now is in the continents. Mohorovičić (1924, 1925, 1927) treated this problem earlier. But not knowing his paper at that time, I performed some simple computations (Escher 1939, 1940a) starting from somewhat different premises. Mohorovičić (1927) supposed that the sialic skin of the earth was, before the separation of the moon, 20 km thick, and that it was afterwards thickened to 40 km by tangential compression of the sima (which he calls sialma). Moreover he supposed that only one third of the sial of the earth was removed to the moon. He took the area of the moon as equal to one fifteenth of the area of the earth and arrived at a thickness of the sial on the moon of $\frac{1}{3}^{5} \times 20 =$ 100 km.

In 1939 I started by the supposition that the sial on earth had not been compressed in tangential direction and that two thirds of the sial layer of the earth were removed to the moon. I accepted 35 km as the thickness of the sial in the continents. The area of the earth being 510×10^6 sq. km, that of the moon 38×10^6 sq. km, they are in proportion of 13.4 to 1. Computed in the manner of Mohorovičić, the thickness of the sial in the moon would be $\frac{2}{3} \times 13.4 \times 35$ km $= 315$ km, whereas I found with a less rough computation, 390 km.

Geophysicists do not agree upon the exact thickness of the sial of the continents. Gutenberg and Richter (1939) give 10 to 30 km as the thickness of the granite layer in the continents.

For our problem, the absolute thickness of the sial on the earth and on the moon is not as important as their comparative thicknesses. According to Mohorovičić the sial on the moon would be 2.5 times as thick as in the continents. According to my view it would be about ten times as thick (Fig. 1).

It does not alter my computation of 1939 in an important way when we take into account the thin sial layer in the Atlantic and in the Indian Oceans. Their area is 155×10^6 sq. km. If we suppose their sialic surface to be 4 km thick, its volume would amount to 620×10^6 cubic km. Then not 11700×10^6 cubic km. but about 11000×10^6 cubic km of sialic material went to the moon. This difference can be neglected for a rough computation. Moreover the areas of the shelves are thus counted double, and it is as continent with a thickness of 35 km and as ocean with one of 4 km, which makes the difference still smaller.

Now we will consider the consequences of the great event for the earth. At the moment of the separation of the moon, there originated a wound in the earth which caused a suction upon the remaining sialic crust of the earth. The suction tore the crust to pieces, which moved towards the wound. Under these sialic floes, there

was viscous magma, already partly solidifying, and beneath that less viscous sial. Behind the floes moving towards the wound, this fluid sial, being less dense than the underlying simatic fluid, came to the surface and formed a thin cover over the sub-

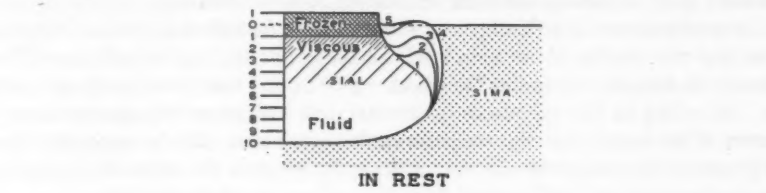

FIGURE 2.—*Hypothetical deformation of crustal layers by flow of sialic blocks*
"Frozen" layer is "Solidified" layer in text.

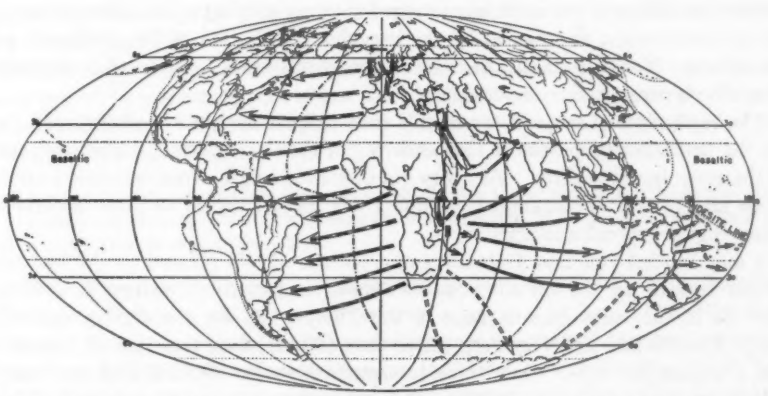

FIGURE 3.—*Flow of sialic blocks toward the Pacific and the European and East-African rift systems*
The arrows show the surfaces travelled by the sial-rafts, now covered by a thin layer of sial.

stratum. At the same time the path of the floes was written on the face of the earth (Figs. 2, 3).

But as the floes moved from all sides towards the wound, there came a moment when they formed a ring and the movement was braked. Perhaps there followed by reaction a small countermovement. The first lay-out of the two Americas, Asia, Australia, Africa, and Antarctica, was born around the quickly solidifying bottom of what would become much later the Pacific Ocean. At the same moment the two main levels on the earth originated, because the lighter sial floated in the denser

sima. The most obvious rent runs between the two Americas on one side and Europe and Africa on the other. If Eurasia and Africa as the biggest continents travelled over a smaller distance than the Americas, it is to be expected that the sial under America is on an average less thick than under Asia and Africa.

In consequence of the thin layer of sial under the Atlantic and Indian Oceans we now find not two but three main levels on the earth: the continental at +875 m, that of the Atlantic and Indian Oceans at −3950 m, and that of the Pacific at −4280 m. According to this hypothesis continental drift took place only once at the very outset of the earth's history, and afterwards never more. At the same time there originated a fracture pattern in the earth's crust, which in the course of the geologic history has again and again pressed its marks on the tectogenetic process.

Vening Meinesz (1934, 1947, 1948) has given a circumstantial article on the formation of a very old shear pattern in the earth's crust as a consequence of a movement of the earth's rotation axis with regard to the crust; and he concludes, among other things, that since this event no great horizontal movements of the continents in respect to one another can have taken place. In the hypothesis here proposed, this old pattern would be a consequence of the separation of the moon from the earth.

This is not the place to enter into a lengthy discussion upon our entirely different points of view. Vening Meinesz started from a pattern on the earth's crust, a minor feature in the face of the earth as compared with my starting point—the asymmetry of the earth's crust as revealed by the irregular distribution of the continents and the oceans. Notwithstanding his refusal of my hypothesis, I still think a reconciliation of both our speculations possible.

The asymmetrical shape of the earth's crust is illustrated in a most pertinent way by the particular properties of the bordering of the Pacific, the place of separation of the moon from the earth. As I may not give a too lengthy consideration, I cannot enter into tectonical contemplations and I will restrict myself to some remarks on seismological and vulcanological facts.

I may remind you that Gutenberg and Richter (1945) pointed to the fact that 80 per cent of the shallow earthquakes appear circumpacific, whereas only 10 per cent lie in the trans-Atlantic zone of the Thetys. As for the earthquakes with deep (300–500 km) and very deep (more than 500 km) foci, these are all located in the circumpacific zone. Only some intermediate shocks (90–250 km) are located outside the bordering of the Pacific. This proves a unique position of the Pacific on the Earth (see also Gutenberg, 1939).

Now I will return to volcanism. James D. Dana in his "Characteristics of volcanoes" (1890) treats almost exclusively of the volcanoes of the Hawaiian Islands. Therefore one could complain of the discrepancy between the title and the contents of the book. The vulcanism of Hawaii is quite different from that of most volcanoes on earth, quite different from that of Vesuvius, Etna, Katmai, Paricutin, Mt. Pelee, and Marapi.

Nevertheless we are glad that the peculiar volcanism of Hawaii has found such an able monographer. As the sima lies at the surface of the Pacific and this is exactly a peculiarity of our planet in contrast to the moon, where the whole surface

must consist of sial, we must consider the basaltic volcanism with its extreme liquidity, low gas pressure, and low content of gas as typical for the earth.

As soon as we have to do with sialic magma the emitted lava is more viscous, which in extreme cases leads to tholoids (Marapi, Mt. Pelee, Tarumai, Usu), whereas the opposite extreme shows great quantities of gas under high pressure, which leads to the formation of calderas. Eruptions, which have created calderas, always eject acid pumice tuffs. (Crater Lake 71 per cent SiO_2, Santorin 65 per cent, Krakatau 66 per cent, Katmai 77 per cent.)

The formation of calderas is a typical expression of avid magma and in its nature something quite different from the volcanic sinks of the Hawaiian volcanoes. The common earthly volcanism ranges between the Strombolian and Perret types of eruption (Escher, 1933a; for diagram see also 1937).

The active volcanism of the earth is, I think, always connected with tension in the earth's crust. This tension is primary or secondary. The former appears in regions of tension (Iceland, African rift valleys), the latter in regions of folding.

A. Holmes (1929) has explained how it is possible to get tension in regions of folding, and I have tried to give an explanation of the connection between this secondary tension and volcanism in a diagram (Escher, 1933b). Whereas the primary tension volcanism is basaltic, the volcanism in regions of folding is intermediate or acid. Primary tension opens the possibility for the sima to escape, whereas by secondary tension in folded regions magma bodies are brought to eruptions which contain acid or mixed material.

Caldera formation is always preceded by a strong eruption, that is an eruption accompanied by great quantities of gas under high pressure. High pressure can only come into existence when the magma body is deep seated. In 1906 the blast of gas of Vesuvius reached a height of 13 km (F. Perret, 1924, p. 45). Perret writes:

> "Let the reader think of globular masses of compressed vapor almost exploding in the air more than 10 km. above the vent, and then try to imagine the original tension of the gas and the degree of its acceleration within the shaft of the volcano."

Now I ask you to try to imagine this tension in the case of the Krakatau eruption of 1883, where the gas and ash column reached a height of at least 50 km (Verbeek, 1885, p. 123). It is obvious that in the latter case the magma body was situated much deeper than in the case of Vesuvius in 1906. Now Rittmann (1933) has explained that the top of the magma body of Vesuvius must lie at 5 km below the sealevel. So I think under Krakatau the top of the magma body must lie a multiple of 5 km below that level.

Here we meet with a problem where we are in need of a physical estimate. We should like to know the functional relation between the depth of the top of the magma body and the height of the gas column.

Nowadays, most volcanologists agree that a caldera is formed by collapse (Williams, 1941; Cotton, 1944, p. 300–316). Instead of the simpler theory of the collapse of the roof of the magma chamber, I have introduced the more complex mechanism of the coring-out of the eruption by the blast of gas, followed by internal sliding (Escher, 1927; 1928; 1929). I prefer this explanation, first because of the

circular form of the real calderas, and secondly because I do not believe that a magma chamber is ever blown empty. It seems more likely to me that as the gas-enriched magma leaves the magma chamber, it is immediately replaced by fresh magma poor in gas. After this fresh magma has cooled down near the walls of the chamber and crystallization has taken place during a certain time, the magma is again enriched with gas. This process goes on till the counter pressure of the roof of the magma chamber is surpassed and a new eruption arises.

Now we will consider the morphology of the moon, which in so many respects differs from that of the earth. Here we must restrict ourselves to the features which Puiseux (1908, p. 121–132) called "*cirques lunaires*," an excellent expression as long as we are not in agreement upon their origin. Goodacre (1931, p. 23–28) calls them craterlike formations and divided them into: large mountain-walled plains, ring-plains, crater rings, and crater pits. Especially in the United States of America the lunar cirques again and again have been explained by the impact of bodies upon the moon. The Meteor Crater of Arizona being a conspicuous phenomenon, it is obvious that the idea of meteors being the bodies of impact was suggested. The newest version of the impact theory is that by Daly (1946) who attributes the moon cirques to the infall of fragments of a planetoid, fragments which were projected beyond Roche's limit after the planetoid struck the liquid earth at or near the equator. Daly proposes mainly three objections against the volcanic origin of the lunar cirques: their size, their high degree of symmetry, and their relative abundance. He asserts that the terrestrial calderas possess a characteristic asymmetry in their ground-plans. Here I cannot follow him, as the typical terrestrial calderas for which I was compelled to devise my caldera theory are to a high degree circular and symmetric.

It is quite true that the diameters of the lunar cirques are tenfold those of the terrestrial calderas. This was for Wegener (1921) a reason not to believe the volcanic origin of lunar cirques and to ascribe them to the infall of meteors, notwithstanding that the Meteor Crater in Arizona measures only one-tenth of the terrestrial calderas, therefore one-hundredth of the lunar cirques.

The hypothesis here developed of the separation of the moon from the earth implies that the volcanism on the moon must have been much stronger than on the earth, as the moon is surrounded by a sialic shell ten times as thick as in the continents (Escher, 1940b). With this both objections of Daly on the size and the abundance of the lunar cirques are refuted. Moreover I feel a difficulty in accepting an impact hypothesis in connection with the mutual intersection of the lunar cirques. On the moon, it is rather the rule than the exception that a circular wall of a greater cirque is interrupted by one of a smaller cirque. A beautiful example is Thebit (diameter 48 km) interrupted by Thebit A (19 km), which in its turn is interrupted by Thebit L (10 km). (*See*, for instance, Goodacre, 1931, p. 169.) It is clear that first Thebit was formed, then Thebit A, and finally Thebit L. On a larger scale we have an example in Stöfler (160 km) the oldest, Faraday (80 km) younger and the cirques A and C (about 30 km) on the ringwall of Faraday the youngest. We find this relationship all over the moon. As an exception, I found only Maurolycus (120 km) which was formed later than the smaller rings on its SE side.

This order of succession from great to small is quite in accordance with a decreasing volcanicity. We know it also on the earth, for instance in the Vulcano Laziale near Rome with the crater lakes of Albano and Nemi. F. E. Wright (1935, 1938) in referring the work of the "Committee on Study of the Surface Features of the Moon" comes to the conclusion that the surface materials of the moon are of the nature of volcanic ash and pumice, high in silica (1935, p. 113); and he says (1938, p. 67) that "If massive rocks occur near the moon's surface they are now covered by pumiceous material." Now that is exactly the material that we know on the earth from calderas and that points to eruptions rich in gas.

Summarizing, I may point to the following combination of problems which are readily explained by the hypothesis here developed on the formation of the moon from the earth.

1. The asymmetrical shape of the face of the earth.
2. The origin of continents and oceans.
3. The formation of a thin sialic layer under the Atlantic and Indian Oceans.
4. The formation of lunar cirques.
5. The much stronger volcanic activity on the moon than on the earth.

I am quite aware that for this hypothesis as well the statement of Daly (1946, p. 118) is valid: "The correlations demand the pyramiding of assumptions." As it explains however so many important problems, I feel authorized to stand up for its rights.

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POST-PLEISTOCENE CONNECTION BETWEEN JAMES BAY AND THE GULF OF SAINT LAWRENCE

BY A. LA ROCQUE

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ABSTRACT

To explain the distribution of certain modern halophytic plants on the shores of James Bay, Potter advanced the idea of a post-Pleistocene marine connection between James Bay and the St. Lawrence drainage; he placed his seaway between Lake Timiskaming and James Bay. Potter's route is considered improbable because of

geological and paleontological objections, which are discussed in detail. Another route, between Lake St. John and James Bay, is advocated and arguments advanced to show that it is not open to the objections against the route chosen by Potter.

INTRODUCTION

Geologic evidence is not confined to facts derived from a study of the rocks, their attitude, their lithologic composition, or the fossils which they may contain. This is especially true of the evidence for the history of the Pleistocene and post-Pleistocene. An interesting example of the unexpected quarters from which geologic evidence may be obtained is a paper by Potter (1932) which seeks to explain the unusual distribution of certain plants by means of a post-Pleistocene connection between James Bay and the St. Lawrence drainage.

Potter's approach to the problem was botanical or more exactly phytogeographic but his conclusions are important to the geologist. His incidental use of evidence drawn from the distribution of the mollusca is of interest; consideration of this aspect of the subject attracted the writer to the problem.

The idea of a connection between James Bay and the St. Lawrence drainage was not new, although Potter seems to have been unaware of the fact. According to Coleman (1922, p. 57), Sir J. William Dawson assumed such a connection to explain most of the glacial phenomena of Ontario and Quebec. Dawson's theory has never found much favor with glacial geologists and had been forgotten long before Potter sought to explain the unusual distribution of his plants. The following paragraph (Coleman, 1922, p. 57) summarizes Dawson's theory and the prevalent opinion of glacial geologists concerning it a few years before Potter attacked the problem from a new approach:

"Dawson explained most of the glacial phenomena of Ontario and Quebec by assuming that the watershed between the St. Lawrence system and Hudson bay was lowered sufficiently to open a broad strait between the Champlain sea and an enlarged Hudson bay or sea. Through this channel he supposed that a powerful current like that off the coast of Labrador swept icebergs and floe ice from the Arctic regions, lowering the temperatures of all eastern Canada. He believed that the boulder clay of the Pleistocene was formed by floating ice and not by a Labrador ice sheet moving toward the south. This view is no longer held by any glacial geologist. . . ."

Dawson's seaway might have remained in the limbo of discarded theories had not Potter sought to explain the distribution of his plants, but the objections which led glacial geologists to discard Dawson's theory are equally valid in disproving Potter's assumed route for a marine connection between James Bay and the Champlain Sea. If Potter's statement on the distribution of his plants is correct, his reasoning justifies a re-examination of the whole question and a careful investigation of the possibility of interpreting the geologic facts to harmonize with the botanical evidence presented by Potter.

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of Zoology, University of Michigan, for information concerning the significance of molluscan assemblages, to Dr. Richard F. Flint for valuable suggestions for the improvement of the manuscript, and to Dr. Nicholas Polunin, Macdonald Professor of Botany, McGill University, Montreal, Canada, for valuable information concerning the plants involved. The maps have been prepared by Mr. Norman J. Willimovsky of the Museum of Zoology, University of Michigan.

POTTER'S ARGUMENT

During the summer of 1929, Potter (1932, p. 70) collected the following species along the shores of Hudson Bay and James Bay: *Zannichellia palustris* L. var. *major* (Boenn.) Koch, *Glaux maritima* L. var. *obtusifolia* Fernald, *Juncus Gerardi* Loisel., *Carex maritima* O. F. Mueller, *Carex norvegica* Willd., *Carex glareosa* var. *amphigena* Fernald, *Plantago juncoides* Lam. var. *decipiens* (Barneoud) Fernald, *Poa eminens* J. S. Presl, *Scirpus rufus* (Hudson) Schrad. These are strict halophytes, that is, plants which will live only near salt water. Potter has given maps showing the distribution of each of these plants in North America. The most important facts about their distribution are:

(1) "In no case have these plants been reported as growing along the shores of Hudson Straits or along the northernmost Labrador coast. Four of them have been rarely found north of Hamilton Inlet; two reach their northern limits in Hamilton Inlet; and five are not known north of the Straits of Belle Isle." (Potter, 1932, p. 71).

(2) The absence of records of these species cannot be explained by the lack of collecting according to Potter, but see my Critique of Potter's argument.

(3) The plants in question are not found inland between James Bay and the Saint Lawrence River. The latter was to be expected, since they are strictly or primarily maritime.

In addition to the strict halophytes enumerated above, Potter also lists the following indifferent halophytes which live in association with the former, but whose distribution can be explained without bringing extraordinary agencies into play: *Potamogeton filiformis* Pers., *Triglochin maritima* L., *T. palustris* L., *Scirpus americanus* Pers., *Juncus balticus* Willd., var. *littoralis* Engelm., *Potentilla Anserina* L., *Myriophyllum exalbescens* Fernald, *Lathyrus maritimus* (L.) Bigel., *Arenaria peploides* L., *Mertensia maritima* (L.) S. F. Gray, *Bidens hyperborea* Greene, an estuarian species, and *Zostera marina* L., a plant confined strictly to salt water.

Potter considers three main explanations of the dispersion of these plants: (1) Driftless areas may have harbored them during the Wisconsin glaciation; (2) dispersal by wind, animals, birds, or water since the recession of the last ice sheet; (3) migration along the shores of a marine connection between the St. Lawrence Basin and James Bay. He discards the first two possibilities for the following reasons:

1. Driftless areas: "Nowhere along the coasts or on the islands of either James Bay or southern Hudson Bay have areas been discovered which escaped the ruthless work of the Wisconsin ice sheet."

2. (a) Wind: "In the case of plants establishing intermediate stations between

two points it is difficult to apply this method of migration to halophytes over areas not favorable to salt-loving plants." (b) Animals other than birds: Fish, amphibia, and reptiles play a very small part in the dissemination of plants. The species in question do not possess seeds so modified as to attach to fur of mammals. "Human travel has not been very great between these two regions and we can scarcely look to this source as the means of introduction. Had these plants any food value or other economic significance, human migrations would of necessity have to be investigated." (c) Birds: Bird authorities agree that birds seldom carry seeds of plants, ". . . and the indisputable evidences of birds carrying seeds either in them or adhering to them mentioned in books evidently apply to birds shot at or not far from, their daily haunts, and not to such as have just made a long journey." (d) Water: The marine waters along the coast of Labrador and Hudson Straits into Hudson Bay were not the route of migration, for, if they had been, the plants should be found in at least a few places along the coast; they are not. Migration via *fresh* water up the streams of the St. Lawrence drainage, over the height of land into the Hudson Bay drainage, is improbable for halophytes.

His only remaining alternative, Potter believes, is migration along the shores of a marine connection between the two bodies of water. He cites the following facts as evidence: (1) The entire area, after being covered with ice, was depressed to a much lower level than at present; (2) The waters of the sea invaded the St. Lawrence Basin at least as far up the Ottawa River as North Bay, the Saguenay River at least as far up as the foot of Lake Saint John; beaches considered to be marine extend this area of marine invasion many miles further inland (*see* Potter's map, 1932, p. 84); (3) Similarly, the land of the Hudson Bay drainage was also depressed and Potter has plotted the known extent of the marine invasion and its probable extent farther inland; (4) The exact limits of the marine invasions, both that from the north and that from the south, have never been accurately mapped, but most geologists agree that they were more extensive than the beaches with marine shells would indicate.

From these facts Potter concludes that a marine connection between Hudson Bay and the Saint Lawrence drainage once existed, and that the halophytes found by him migrated along this route from the Saint Lawrence to Hudson Bay. His conclusions are quoted in part (Potter, 1932, p. 83):

"1. A marine invasion occurred in the St. Lawrence Basin after the recession of the Wisconsin Ice Sheet, which probably extended north to include the Lake Temiskaming region.

"2. A similar invasion of the sea occurred in the Hudson Bay region contemporaneous with that of the St. Lawrence Basin, which extended southward to within a few miles of the height of land.

"3. Evidence seems to indicate that a possible marine connection existed between these two known submergences involving the Lake Abitibi area.

"6. The above marine connection (if it existed) would have offered suitable conditions for the migration of halophytic plants along its shores.

"7. Undoubtedly the northern margin of the Champlain Sea and the southern margin of the Hudson Bay inundation were relatively close and, even though a marine connection may have been lacking, the factors effecting local plant-distribution may have been sufficient to have bridged the slight gap and thus account for the occurrence of the plants in question in the region of Hudson Bay."

CRITIQUE OF POTTER'S ARGUMENT

It should be stated at the outset that Potter's reasoning is sound and his conclusions justified in part providing his presentation of the distribution of the species in

question is correct. Potter's whole argument rests on two assumptions, first, that the halophytes discussed are not continuously distributed from Hamilton Inlet around the Ungava peninsula into Hudson Bay, and, second, that these species could not spread overland or along bodies of fresh water. If these two premises are accepted, his explanation (1932, p. 70) is highly plausible.

The writer agrees with Potter in discarding possibilities (1), (2a, b, and c) for the following reasons:

1. Driftless areas: Potter's statement on the absence of driftless areas in the region in question is strengthened by the Glacial Map of North America (Flint *et al.*, 1945).

2. (a) Potter might have added that the seeds of halophytes are not built so as to be carried easily by wind; they have nothing like the appendages of seeds of dandelion and milkweed which permit wide distribution of these weeds.

2. (b) Potter seems to have neglected the possibility that seeds might be eaten by mammals, pass through their digestive tract, and emerge without losing their viability. This possibility is disposed of on two counts: first their being eaten by mammals is unlikely because of their unattractiveness as food and the fact that they are most abundant in water, far from the usual habitat of the seed-eating mammals; secondly, if they should be eaten by mammals, it would be most unlikely that they should be retained within the mammals' body for the several days that would be required for the mammals to travel from the shores of the St. Lawrence to those of Hudson and James Bays.

2. (c) The same possibility appears more likely in the case of migratory birds, since their speed is greater and they cover greater distances in their annual migrations. On the other hand, ornithologists are generally agreed that birds do not carry seeds far, as Potter points out.

2. (d) Migration via fresh water up the streams of the St. Lawrence drainage over the height of land into the Hudson Bay drainage appears very improbable. The seeds would have to drift upstream in the first part of their journey and retain their viability for many months.

There remains the possibility of natural dispersal by gradual spread along the coasts of Labrador, Ungava, and Hudson Bay. Potter discards this possibility on the grounds that if it were valid the plants should be found in at least a few places along the coast; he believed that they were not. Other botanists disagree with Potter on the distribution of his critical species. Polunin (1940, p. 40) has summarized their views as follows:

> "... it must be mentioned that not merely *Zostera* but also several others (most notably "Carex maritima", "Carex glareosa var. amphigena", and "Carex norvegica") of the eleven plants that he mentions in support of his view, are now proving to be much more widely distributed in the north than had been thought, and may easily have migrated by the present-day sea connection through Hudson Strait, where indeed at least two of them now grow plentifully."

The extended distribution of the species mentioned by Potter is detailed in papers by Porsild (1932) and Polunin (1940; 1943a; 1943b). This weakens Potter's argument considerably, but it does not completely exclude the possibility of a Pleistocene connection between James Bay and the St. Lawrence drainage.

If a post-Pleistocene marine connection between the Saint Lawrence drainage

and James Bay is not impossible, its probable location is a point worth discussing. Potter considered only one of the possible routes, although his map (1932, p. 84) shows at least two long bays in the Champlain Sea which could quite conceivably have served the purpose as well as the Lake Timiskaming route, namely the Gatineau-Lièvre valley and the Saguenay-Lake St. John valley. The pros and cons for each route should be examined carefully in order to decide, if possible, which of the three is the likelier route.

As it stands now, Potter's hypothesis is open, on geological and paleontological grounds, to objections so serious that were it not possible to argue for another route, one would be forced to abandon the hypothesis altogether. These objections apply only if the marine connection is assumed to have been located in the area between Lake Timiskaming and James Bay or the Gatineau-Lièvre arm of the Champlain Sea.

GEOLOGIC OBJECTIONS TO THE WESTERN ROUTES

PATH OF RECESSION OF THE WISCONSIN ICE SHEET

Daly (1934, p. 101, fig. 57) and Flint *et al.* (1945) have summarized the information available on the position of the front of the Wisconsin ice sheet as indicated by moraines. The moraines in the area between the Great Lakes and James Bay run roughly NNW—ESE. From these maps alone it is obvious that by the time James Bay was free of ice, the ice had been gone from the headwaters of the Ottawa River for a long time; as long as, and probably longer than, the time between the birth of Lake Algonquin and its termination. This time was surely sufficient for a certain amount of plastic recoil in the Lake Timiskaming region, very probably enough to initiate the recession of the Champlain Sea from the Ottawa valley.

LAKE BARLOW-OJIBWAY

. The formation, growth, and draining of Lake Barlow-Ojibway have been discussed by Coleman (1909, p. 284–293), Wilson (1918), and Antevs (1925, p. 75). Norman (1938, 1939) added supplementary data on the extent of the lake eastward. Its full extent is shown on the Glacial Map of North America (Flint *et al.*, 1945). Lake Barlow-Ojibway appears to have been a body of fresh water. It was not drained until the ice front in Manitoba receded to the coast of Hudson Bay in the vicinity of York Factory (Figs. 2, 3); at this time, only the western shores of Hudson Bay above York Factory had been freed of ice, and another long period of time elapsed before the ice front receded seaward from the western coast of James Bay. To accept the Lake Timiskaming connection, one would have to assume (1) that Lake Barlow-Ojibway was a body of salt water, since it would have been connected with the Ottawa arm of the Champlain Sea, (2) that the connection was first established with Hudson Bay, and (3) that it remained in effect for a very long period of time while the ice retreated at least to the western shore of James Bay. Potter's halophytes might have penetrated to Hudson Bay by this route and subsequently migrated south and east along the coast. In that case they should be found far up the southwestern shore of Hudson Bay, and *Carex norvegica* and *Scirpus rufus* are (Potter, 1932, maps

1, 6). Porsild (1932, p. 90–94) records *Zostera marina* from Cape Eskimo (Latitude 61°) and thinks that "Further investigations will probably show that *Zostera* occurs extensively in sheltered places in the southern part of Hudson Bay." He also records *Zannichellia palustris* from Cape Henrietta Maria. The other five of Potter's crucial species are absent from the region west of Cape Henrietta Maria.

On the other hand, Coleman (1922, p. 44) categorically asserts that the waters of Lake Barlow-Ojibway were not salt, that the North Bay outlet of the Lake

> "must have opened considerably later than the highest marine stage at Ottawa, when elevation of the land had proceeded to the extent of more than 140 feet; and we must assume that ice still occupied the Nipissing-Timiskaming region when the highest marine level occurred at Ottawa."

Subsequent workers agree with this opinion (Flint *et al.*, 1945).

Moreover, the occurrence of the halophytes along the west coast of Hudson Bay can be explained by normal extension of their range from the James Bay region *after* their introduction into the James Bay area.

CHRONOLOGY OF UPLIFT

It has been established that there is a considerable lag between removal of the ice load from a glaciated area and the beginning of uplift; and that the uplift is gradual and follows approximately the path followed by the receding ice front. Unfortunately, there is no precise chronology for the recession of the ice from the region south of the Great Lakes to the final disappearance of the ice from the area of Hudson Bay. Antevs' measurements (Daly, 1934, p. 86, fig. 46) contain three gaps which Antevs represents by x, y, and z years. Nevertheless, 13,000 years elapsed between the time at which the ice front stood at Stony Lake, Ontario, and that at which it stood near Mattawa, Quebec; and 2,000 years between its position at Montreal River and Cochrane, Ontario.

No way has yet been found to measure the interval between recession of the ice front and the beginning of uplift. Still, some approximations based on indirect evidence provide interesting data for speculation. Coleman (1922, p. 71) has summarized the succession of events in a table which strongly suggests that by the time the ice front had receded sufficiently to drain Lake Barlow-Ojibway into the western part of Hudson Bay, the uplift of the region between Lake Timiskaming and James Bay had proceeded sufficiently to establish the height of land and to prevent a connection of Hudson Bay and the Ottawa arm of the Champlain Sea.

CONSEQUENCES OF THE ASSUMED CONNECTION

The marine connection between James Bay and the Ottawa arm of the Champlain Sea, advocated by Potter, would make it necessary to suppose (1) that uplift in the area of the present height of land between Lake Timiskaming and James Bay was delayed at least until Lake Barlow-Ojibway was drained by the retreat of the ice front across the western shore of Hudson Bay in the vicinity of York Factory; (2) that uplift in the area from Lake Timiskaming down the Ottawa River to Montreal and down the Saint Lawrence to Quebec and beyond was not sufficient to cause the sea to withdraw from the upper reaches of the Ottawa. Granting these two assump-

tions, the expectable consequences in the Great Lakes area are entirely out of harmony with the known facts. Failure of the crust to rise would have resulted in an invasion of the sea into the area known to have been covered by the Algonquin Great Lakes (second stage) and the Nipissing Great Lakes. Fairchild (1907; 1919) has pointed out that the Champlain Sea probably extended much farther into the Lake Ontario basin than previously supposed, but his arguments could never be stretched to include such a large area as that required to preserve the connection predicated by Potter.

Coleman (1922, p. 54, fig. 12) and Flint *et al.* (1945) have mapped the maximum marine invasion in Ontario which seems to be the greatest extent which can be ascribed to the Champlain Sea on the geological evidence available.

Such an extension would explain the whale remains in Michigan and the supposed survival and adaptation to fresh water conditions of certain invertebrates now living in the depths of the Great Lakes, but both of these may be explained in other ways. In addition, evidence of uplift in the Ottawa Valley is indisputable and it is established that it was not abnormally delayed.

BOUNDARIES OF THE CHAMPLAIN SEA

Potter was careful to distinguish between beaches of undoubted marine character, as attested by the presence of marine fossils, and the probably marine ones whose character is not proven by the presence of marine fossils. His map (1932, p. 84) extends the area of "marine submergence based upon fossils" 2 whole degrees of longitude farther west than Lake Coulonge which, according to Potter and other geologists, is the westernmost locality at which marine fossils have been found in the valley of the Ottawa River. Johnston (1916, p. 4), thoroughly familiar with the Pleistocene and Recent deposits of the Ottawa Valley, says:

> "The sea extended far up the Ottawa valley, possibly as far as the head of Lake Timiskaming. Marine fossils are known to occur in the clays as far west as Lake Coulonge, about 70 miles west of Ottawa. Above this point the Ottawa valley is narrow, so that during the time of marine submergence the upper part of the valley was a long, narrow estuary."

Thus, it seems that Potter was alone in supposing that the Champlain Sea extended up the Ottawa River farther than the head of Lake Timiskaming, excepting the early hypothesis of Dawson.

Potter attaches great weight to a statement by Bell (1896) that ". . . in James Bay today no living marine mollusks are to be found, except perhaps in the northern part, owing probably to the muddy and brackish nature of the water." and surmises that "A similar condition might well have existed during the Champlain Submergence in which the waters were muddy and brackish, not permitting the existence of marine life, yet allowing the existence of plants of maritime habitats along its shores."

Richards (1936, p. 534) found living specimens of *Littorina rudis* (Maton), *Macoma balthica* (L.), *Mytilus edulis* L., and *Acmaea testudinalis* (Müller) at Charlton Island in the southern part of the bay so that Bell's statement must be heavily discounted as based on insufficient observation.

It is not likely that a body of salt water extensive enough to connect James Bay with the Saint Lawrence drainage would long remain without molluscan inhabitants,

and the abundance of such forms as *Saxicava arctica* (L.) and *Macoma balthica* (L.) elsewhere in the Champlain Sea is proof enough of the spread of marine mollusks. A newly established body of salt water would remain for a time without a molluscan fauna, but it is unlikely that their spread would be slower than that of Potter's halophytes; if anything it would be quicker, for their larvae are carried long distances by marine currents, which, Potter emphasizes, are not important in the distribution of the halophytes.

These facts considerably weaken Potter's argument for the Lake Timiskaming seaway, as the beaches in the area above Lake Timiskaming cannot all be assumed to be marine. Possibly some are, but the absence of marine fossils makes it difficult to distinguish them from the beaches formed by Lake Barlow-Ojibway.

PALEONTOLOGICAL OBJECTIONS TO THE WESTERN ROUTES

CHARACTER OF THE PLEISTOCENE FAUNA

The presence or absence of certain fossils and the comparative abundance or scarcity of those found furnish useful information on the nature of the environment when the marine fauna was living. Since the species of the Champlain Sea are generically and sometimes even specifically identical with the species now living in the Atlantic, useful information can be derived from direct comparison with the living species. Also, the order in which the marine fauna invades an area newly covered with salt water sheds some light on the manner in which the Champlain Sea was populated.

Fortunately, Sir J. William Dawson has thoroughly investigated the fauna of the Pleistocene in the Saint Lawrence basin. Extensive lists of species given in his "Canadian Ice Age" (1893) have been brought up to date and arranged in tabular form by Goldring (1922, p. 160–164). The writer had the privilege of examining many of Dawson's localities in the Saint Lawrence Valley during the field season of 1936, and of collecting a large number of fossils which are now deposited in the collections of the Geological Survey at Ottawa. A striking feature of the various faunules is the great variety of species which occur in the Pleistocene clays all the way up the St. Lawrence valley from Rivière-du-Loup to Montreal and the sharp drop in number of species in deposits above Montreal. This fact is clearly brought out by an examination of the tables in Goldring (1922).

INVASION OF NEW MARINE AREAS

A recent paper by W. S. S. van Benthem Jutting (1946) gives details of the invasion of the marine fauna on the island of Walcheren, province of Zeeland, Netherlands. The dikes of the island were bombed by the Allies in October, 1944, to hamper the German defence of the island. After the surrender of the German forces, the industrious Dutch repaired the dikes and pumped the water from the island. Mrs. van Benthem Jutting, realizing the exceptional opportunity offered her, studied the effects of the flooding and recorded the species which were able to establish themselves. Thus, the duration of the invasion, October 1944 to October 1945, and the fauna which established itself during that time are known. Those species which

established themselves on Walcheren are those which may reasonably be expected in the vanguard of an advancing sea, those quickest to populate new areas of salt water. Mrs. van Benthem Jutting's list is of exceptional interest for the problem under study: *Mytilus edulis*, very abundant; *Membranipora* (*Electra*) *crustulenta* (Pallas) var. *fossaria* (Hincks), a bryozoan, also very abundant; *Balanus crenatus* Bruguière "another settler, even the most rapid of them all"; *Ostrea edulis* L. "much less frequent than any of the preceding species"; *Cardium edule* L. "a shovelful" and *Mya arenaria* L. "about a dozen"; *Carcinides maenas* L. "two carapaces"; *Asterias rubens* L. "observed only once"; "the common periwinkle (*Littorina littorea* L.) has not been recorded."

Mrs. van Benthem Jutting concludes

> "The marine phase in the history of Walcheren has not led to a condition in which the fauna of the flooded area was representative of the sea fauna in the immediate vicinity of the island, i.e. the fauna of the Oosterschelde, the Westerschelde or the North Sea off Westkapelle and Domburg. For that purpose the time during which the invasion lasted was far too short and the communication between the inland sea and the exterior too fragmentary, so that only a fraction of the number of species had the opportunity to enter. The marine fauna of Walcheren therefore was composed of species which either could swim actively or which had pelagic larvae. Animals with a slow rate of locomotion such as various gastropods, had little or no chance to take possession of the new habitat." (1946, p. 7).

Her list shows two species in common with those found above Montreal, viz. *Mytilus edulis* and *Balanus crenatus*. The overwhelming majority of fossils present in the higher beaches above Montreal are *Saxicava arctica* L., *Macoma balthica* L., *Mytilus edulis* L., *Cylichna alba* and *Balanus crenatus* Brug. In at least one deposit near Chelsea, Quebec, some 10 miles up the Gatineau Valley from Ottawa, these species occur in great abundance in a layer some 10 inches thick. In this deposit, the valves of the pelecypods and the plates of the barnacles are dissociated and the *Mytilus* occurs only as fragments. This is interpreted as evidence of sorting and concentration at a level lower than the original levels of accumulation. It may be assumed that the Champlain Sea invasion above Montreal was of shorter duration than at Montreal and below it. Otherwise the abundant fauna preserved at Montreal would have had time to spread to Ottawa and perhaps beyond.

The succession of events in an area newly invaded by the sea after the retreat of ice provides interesting data for estimating the boundaries of the Champlain Sea. Three general principles influence the advance and retreat of the marine fauna: (1) The maximum extent of the sea will be reached as soon as the ice has retreated from a given area; the highest beaches will form first. (2) After an undetermined length of time the land will rise slowly, the first or highest beaches will emerge and be exposed to erosion so that any fossils in the higher deposits will be more likely to be destroyed by erosion than those in succeeding beaches; likewise, the possibility of their being washed down to lower beaches is greater. (3) The marine fauna will not establish itself immediately. The upper reaches of an arm of the sea as large as the Ottawa-Saint Lawrence embayment will be brackish or even quite fresh for a long time because of the melt-water emptying into it from the retreating glacier.

It follows that the establishment of marine species will depend on three factors: (1) the balance between increasing salinity and the tolerance of a given species to brackish-water conditions, (2) the method of dispersal of a given species; the first

to invade will be free-swimming forms and those with pelagic larvae that can be carried by currents. Of these, the first to appear will be those which can tolerate brackish-water conditions, such as *Mytilus edulis* and *Saxicava arctica*, then those that require an increasingly greater salinity. Last to invade will be the viviparous species and those that lay eggs attached to the substrate. (3) The land will rise and the sea retreat first in those areas first uncovered by ice. If the river valleys had not antedated the invasion of ice, the marine invasion would have been very brief and ephemeral salt lakes would have been formed by marine waters trapped in the upper reaches of the arms of the sea.

Applying these principles to the Ottawa-Saint Lawrence Valley, one might expect the following succession of events: *Stage 1*: Gradual freeing of the lower Saint Lawrence from the mouth of the Ottawa River at Montreal downstream to the Gulf of Saint Lawrence (Fig. 1). This would correspond roughly to Coleman's early (low) marine stage (Coleman, 1922, p. 71). At this stage, the euryhaline and fast-spreading species advanced up the Saint Lawrence Valley as far as Montreal. *Stage 2*: Maximum marine stage of Coleman. Note that Coleman does not consider the possibility of the sea having reached into the area of Lake Ojibway. He seems to think that the height of land between Lake Timiskaming and James Bay had been sufficiently elevated to prevent the marine invasion from extending farther north. At this stage, euryhaline and fast-spreading species followed the sea and the retreating ice front up the Ottawa Valley as far as Lake Timiskaming. Species that spread more slowly reached the Montreal area. *Stage 3*: The sea had begun to retreat down the Ottawa River. The land had started to rise in the Ottawa Valley region, some of the higher beaches were emergent and were being attacked by erosion. *Stage 4*: The sea had retreated still farther down the Ottawa and the area was being slowly elevated. The Saint Lawrence Valley in the Montreal area and below it had been covered by the sea since the end of stage 1; therefore it is not surprising to find that the fauna of the marine beds there contains both slow and fast-spreading species with an abundance of genera, species, and individuals.

LAKE SAINT JOHN ROUTE

GENERAL DISCUSSION

Potter's map (1932, p. 84) so obviously suggests another possible marine connection between James Bay and the Saint Lawrence River that it is surprising that he did not discuss it at all; this is the route between Lake Saint John and James Bay. Potter's arguments are equally valid for this route. He shows a beach with marine shells at 300 feet above sea level on Ha Ha Bay and a "probable marine submergence based upon raised beaches" all around Lake Saint John. His probable marine submergence in the James Bay area extends just beyond the eastern shore of Lake Waswanipi. He may have considered this possibility and discarded it because of the greater elevation of land between his two supposed marine submergences. Lake Askitichi, at the headwaters of the south branch of the Ashwapmuchwan River has an elevation of 1,288 feet above sea level and Lake Obatogamau, at the headwaters of the Etchipotchi River, stands at 1,247 feet above sea level; thus the height of land

between these two lakes is at least more than 1,288 feet, whereas the elevation of Lake Saint John is not more than 341 feet and the highest marine beach recorded by Potter is 1,000 feet, at the western end of Lake Saint John. Nevertheless, accepting his data as correct, it is not much more difficult to assume a depression of 300 feet in

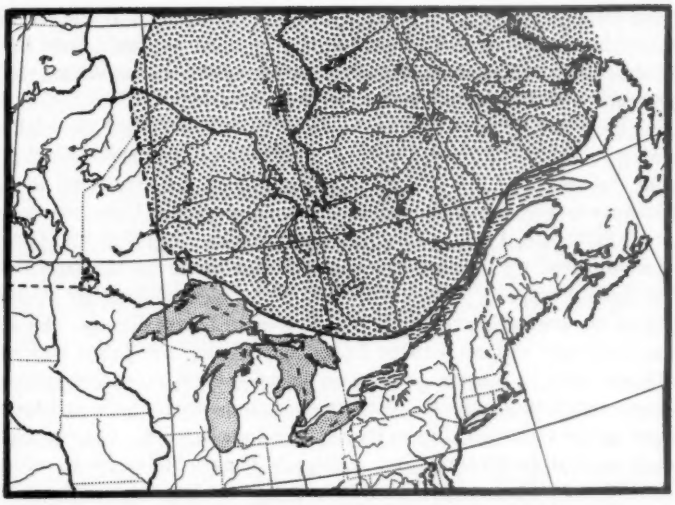

FIGURE 1.—*Early (Low) marine stage in eastern Ontario*

Only the features discussed in this paper are shown. (After Coleman, 1922) Pattern: ice, coarse stippling; fresh water, fine stippling; marine invasions, short dashes.

the divide area northwest of Lake Saint John than one of 85 feet north of Lake Timiskaming. Daly (1934, p. 106, fig. 64) shows the 250 meter isobase for the post-Glacial marine limit passing just northwest of Lake Saint John, which would indicate ample depression in this area to submerge the height of land. Moreover, Daly (1934, p. 107) points out that the isobases are less than the actual upheaval, which was probably at least 50 per cent greater.

If the Lake Saint John seaway is to receive serious consideration, it must not be open to the objections which have been raised against the Lake Timiskaming route. It will be necessary, therefore, to consider each of these in detail before its acceptance can be considered.

PATH OF RECESSION OF THE WISCONSIN ICE SHEET

According to Antevs (1925, p. 78) "the ice edge can be reconstructed by drawing lines at right angles to the striae, because the majority of the preserved scratches were formed near the ice edge shortly before the ultimate uncovering. . . ." Applying this principle to Antevs' Figure 27, which shows glacial striae in eastern Canada and the probable positions of the ice edge at various stages, and the Glacial Map of North America (Flint *et al.*, 1945), it will be seen that the ice front in the area between

Lake Saint John and James Bay retreated roughly southwest to northeast. Thus, Lake Saint John and the area between it and James Bay were evidently freed of ice at approximately the same time, and relative depression would be about the same along a line connecting Lake Saint John with the mouth of the Nottaway River.

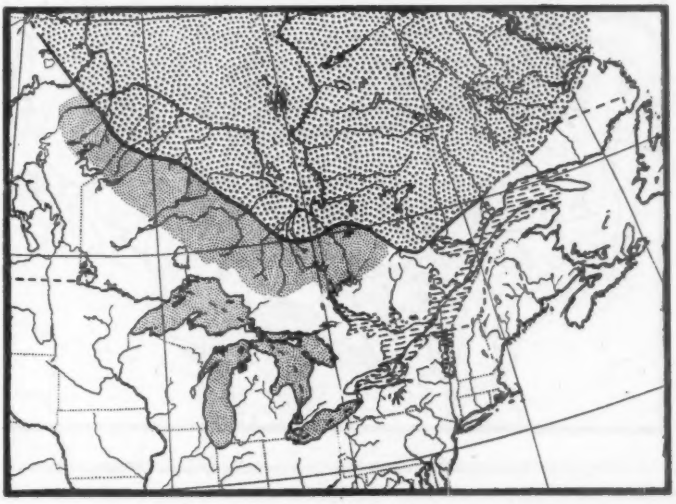

FIGURE 2.—*Maximum marine stage in eastern Ontario*

Lake Ojibway, Gilbert Gulf, Mattawa Outlet of Nipissing Great Lakes. (After Coleman (1922), Daly (1934), Norman (1938).) Pattern: ice, coarse stippling; fresh water, fine stippling; marine invasions, short dashes.

Information on glacial striae is not as extensive as one would wish. Still, Norman (1938, Fig. 2) has shown the annual moraines and the approximate position of the ice front at assumed 100-year intervals for the Chibougamau area and these show the same direction for the recession of the ice front.

The pattern of glacial striae (Antevs, 1925, Fig. 27) and the Glacial Map of North America (Flint *et al.*, 1945) strongly suggest retreat in the form of two lobes, one in Ontario, the other in Quebec. This is also suggested by Norman's map (1938, Fig. 1) but not by Daly's (1934, p. 101, fig. 57). On the whole, Antevs', Flint's, and Norman's maps are probably more correct than Daly's especially since Norman's is based on detailed field work in the Chibougamau area. The marine invasion, then, would have to wait until James Bay had been uncovered, at least in part, and the ice front had retreated at least to the line of the Rupert or Eastmain Rivers (Fig. 4).

LAKE BARLOW-OJIBWAY

From the arguments just advanced, it follows that the marine invasion must have come after the draining of Lake Barlow-Ojibway. Uplift in the Lake Timiskaming area would have proceeded for quite a long time, i.e. during the existence of Lake Barlow-Ojibway, plus the time necessary to free the western shore of James Bay of

ice (Figs. 2, 3). Therefore, the very arguments which militate against the Lake Timiskaming connection are favorable to the Lake Saint John route.

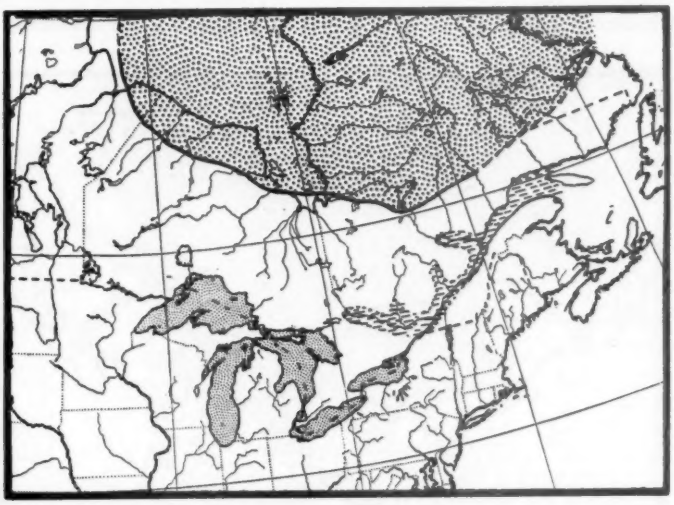

FIGURE 3.—*Conditions shortly after the draining of Lake Ojibway.*

Note that James Bay is still covered with ice. There has been no change in the Lake Saint John area because the ice has retreated from it much later than from the Ottawa Valley. Pattern: ice, coarse stippling; fresh water, fine stippling; marine invasions, short dashes.

CHRONOLOGY OF UPLIFT

The uplift in the Lake Timiskaming area, whatever its extent, is not involved in the consideration of the Lake Saint John connection. However much the Ottawa and Upper Saint Lawrence valleys may have been uplifted, the Saguenay River empties into the St. Lawrence far below the present salt-water line. The writer collected living marine mollusca at Saint-Roch-des-Aulnaies, L'Islet County, Quebec, in 1936. Thus, the margin between the mouth of the Saguenay River and the beginning of brackish or fresh water is ample to permit a marine fauna to invade the Saguenay Valley. In addition, Tolmachoff's (1927) record of marine pelecypods at the foot of Lake Saint John is confirmation, if needed.

CONSEQUENCES OF THE ASSUMED CONNECTION

Assuming a marine connection between Lake Saint John and James Bay does not influence in any way the events in the area west and south of Lake Timiskaming. It is necessary to suppose that the marine connection came into existence after Lake Barlow-Ojibway was drained, for at the time of the draining of the lake James Bay was certainly still covered with ice.

BOUNDARIES OF THE CHAMPLAIN SEA

Very little attention has been paid to the boundaries of the Champlain Sea in the Lake Saint John area. Most of the published maps stop in the vicinity of Quebec

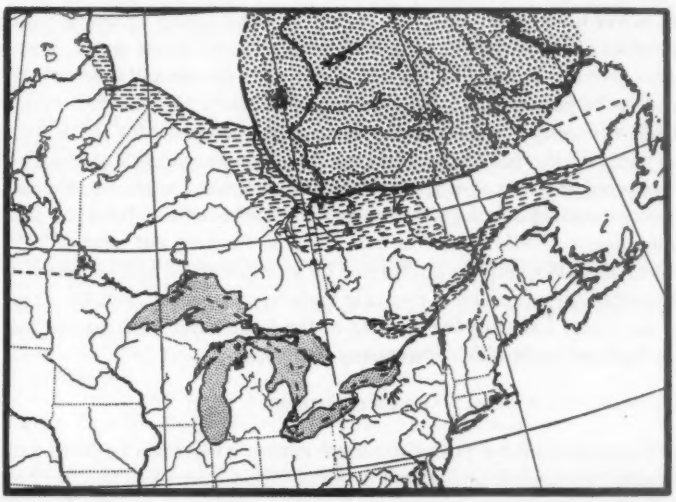

FIGURE 4.—*Possible marine connection between Lake Saint John and James Bay when the ice front had retreated from part of James Bay.*

Note the extension along the coast of Hudson Bay. This area, having been recently freed of ice, would still be depressed. Pattern: ice, coarse stippling; fresh water, fine stippling; marine invasions, short dashes.

or even farther up the Saint Lawrence (Daly, 1934, p. 103, fig. 58; Coleman, 1922, p. 53, fig. 12; Radforth, 1944, p. 21, fig. 14; Goldring, 1922, p. 180, map 2). Tolmachoff (1927, p. 123–125) reports *Saxicava arctica* L. and *Macoma balthica* L. in Champlain Sea deposits 255 to 275 feet above sea level "about an eighth of a mile above the earth fill dam on the west branch of the Grand Discharge . . ., the main channel of the Saguenay River into which Lake Saint John empties its waters." So far as the writer knows, no find of Pleistocene shells has been recorded farther west and north toward the height of land in this area, but there is a remote possibility that some such record may be found in the many reports which make incidental mention of glacial and recent geology in the area. The boundaries of the marine invasion around Lake St. John, so far as they are certainly known at present, may be taken as those shown on the Glacial Map of North America (Flint *et al.*, 1945).

PALEONTOLOGICAL OBJECTIONS

The absence of fossils in the gap between Lake St. John and James Bay is just as serious an objection to the Lake St. John connection as to the Lake Timiskaming one. But it is only one objection, not one of many. Moreover, it is an objection which may be removed by more intensive search in the area, still difficult of access and far from thoroughly explored.

INVASION OF NEW MARINE AREAS

This objection is much less weighty against the Lake Saint John connection than against the Lake Timiskaming one. The length of time necessary to establish a marine fauna by the Lake Saint John route, if it can be measured by the distance involved, is five to ten times shorter. Moreover, the drainage involved, which might be expected to cause a certain amount of freshening of the sea water, is much less. This would give the marine fauna a better chance to invade and become established. Incidentally, the path of the halophytes is much shorter also, so that a marine connection was necessary for a shorter time, i.e. that necessary for the halophytes to cross the rising divide, before the sea began to recede on each side as the divide rose.

Positive proof or disproof of the Lake Saint John route might be revealed by examination of the surface geology of the area between Lake Saint John and James Bay. A certain amount of field work has been done in the area but most of it is purely descriptive; the data should be interpreted by an experienced glacial geologist.

The hypothetical succession of events in Ontario and Quebec from the Trent Outlet Stage of the Great Lakes to the opening of the Lake Saint John connection is presented in the four maps which accompany this paper.

CONCLUSIONS

Potter's arguments for a post-Pleistocene marine connection between James Bay and the Saint Lawrence drainage are worthy of consideration as a working hypothesis in investigating post-Pleistocene geology in the James Bay area. The Lake Timiskaming route which he chose as the marine connection is improbable on geological and paleontological grounds. Of the two other possible locations, the Lake Saint John route appears to be the more probable because it is not open to the objections raised against the Lake Timiskaming route. Tangible proof of the existence of the Lake Saint John connection must await positive evidence which might be provided by field work in the Lake Saint John-James Bay area.

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